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REFERENCE

Lane, B. H., 1935. Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C., 25c.)

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GENETIC RELATIONS OF CAVES TO PENEPLAINS AND BIG SPRINGS IN THE OZARKS

J HARLEN BRETZ

ABSTRACT. 1. Most Ozark caves were made by a phreatic circulation of meteoric water beneath the mature topography which preceded peneplanation of the Ozark dome.

2. During old age, cave-making ceased as hydrostatic head disappeared with reduction of the uplands. To a large extent the existing caves then became filled with residual red clay filtered down from the peneplain soil. Remnants of this clay fill are considered as valid a record of peneplanation as are the flattish uplands beveling structure.

3. Later uplift caused deep dissection of the peneplained dome. This lowered the water table and brought most of the caves into the vadose zone.

4. The ensuing (present) vadose circulation, by removing the clay fill, is devoted more largely to restoring the original phreatic caves than it is to enlarging them.

5. The big springs of the Ozarks are interpreted as discharges from members of the early cave system which lay so low that post-uplift dissection has just reached them. Some of these subterranean aquifers are still operating under phreatic conditions.

INTRODUCTION

THE concept of solutional cave-making largely by a circulation in the phreatic zone (below the water table) should be credited to W. M. Davis. He became interested in cave origins too late in life to undertake subterranean explorations and developed his theory (1930) entirely from a literature containing almost no suggestion of the origin he proposed. There are few comparable cases where the significance of field facts has been almost completely missed and only vague interpretations made on an assumption. In this case, the assumption was that vadose water, descending to and perhaps flowing along the water table, was responsible for solutional caves in limestone and dolomite. Critically analyzing the data he found,

Davis pointed out many features of cave patterns, proportions and relations which could not be satisfactorily explained as consequent on vadose circulation alone but had a logical explanation in the theory of subwater table origin.

The clash of rival theories is a healthy state because new field criteria thus are sure to come to light. But the cave problem is complicated because every enterable cave must have had some experience with vadose water, even if its origin dates back to completely saturated conditions. Cave features resulting from the earlier experience may thus become obscured or obliterated during the later one. More critical study of caves obviously was the answer to the Davis challenge.

Such studies by Swinnerton (1932), Stone (1932), Gardner (1935), McGill (1935), Malott (1936), and Tullis and Gries (undated) have specifically considered caves in various limited regions and have reached different conclusions. Caves vary in character in terms of differing calcareous formations containing them and of differing character and history of the topography and drainage above them. An outstanding example of refusal to follow Davis was Malott's "invasion theory." The Indiana caves of his studies were interpreted as the product of lost rivers, sinking creeks and sinkhole drainage of storm water, all discharging into "primitive" phreatic routes and vastly enlarging them to the dimensions of Marengo, Wyandotte and other capacious caves. In this interpretation, Malott was right. The vadose enlargements in Indiana caves have almost destroyed the relatively small phreatic caves of earlier date, only small remnants recording his "primitive tubes."

On the other hand, Tullis and Gries found the large group of Black Hills caves to conform to the Davis phreatic theory and to carry almost no record of vadose enlargement. Stone, studying Pennsylvania caves, also found convincing evidence of phreatic origin.

From a study of more than a hundred caves in 16 states, including Indiana, South Dakota and Pennsylvania, the writer stated in a paper (1942) postdating publication by all the above-mentioned authors, that 44 of the caves he had examined were entirely phreatic in origin, 49 carried a record of vadose enlargement of phreatic cavities and 3 were wholly vadose in origin. He later undertook a detailed study of the caves of southern Missouri for the Missouri Geological Survey, on the

results of which this paper is based. Because most of the 138 Missouri caves he studied were largely of phreatic origin, portions of the manuscript for the Missouri Survey were sent to Malott for comment and criticism. In the ensuing correspondence, Malott indicated surprise at the extensive cave development in the Ozarks where no lost rivers, only a few sinking creeks and but little karst topography exist. He also stated that he had never seen the red clay fills or the spongework which the writer had argued can exist only in cavities of phreatic origin. All this is understandable in the light of his experience; limited to caves whose extensive vadose chambers have been greatly enlarged from small phreatic predecessors. The Black Hills caves, with no vadose record, might have puzzled him still more.

In sum, most subterranean drainage ways in calcareous rock conform to regional habits just as do surface drainage ways, and in addition many record experience in varying degree with the two contrasted environments, the phreatic experience preceding the vadose.

ACKNOWLEDGMENTS

Several months of cavern study under the Missouri Geological Survey and Water Resources has more than doubled the writer's list of caves studied and has enlarged his understanding of the relations of Ozark caves to the overlying topography. Thanks are due to the State Geologist, Dr. E. L. Clark, for permission to publish this paper in advance of the detailed report the Missouri Survey will issue.

OZARK PENEPLANATION

"The Ozark dome" is a widely used phrase for a far from simple structure constituting the southern half of Missouri and overlapping into northern Arkansas. Its complexities include faulting and folding under different stresses at different times. The net result, under later erosion, is an asymmetrical domelike area, a topographic high of the present drainage with the subcentral Precambrian St. Francis Mountains surrounded by irregular belts of Paleozoic sediments ranging from Cambrian to Pennsylvanian in age and dipping radially outward a little more than does the surface. Varying degrees of resistance to erosion in rocks of this belted pattern have produced a series of cuestas whose scarps face the central area.

At least three partially base-leveled surfaces are recorded in the present topography. The cuestas have existed since the second cycle of erosion was initiated and their crests and upper back slopes carry some of the best preserved records of the first two base levelings.

Previous interpretations of Ozark physiography have called for two erosion cycles preceding the present, but not in all cases the same two. Hershey (1895), the first to announce the idea of multiple cycles for the Ozarks, found his upper surface on the broad upland underlain by Mississippian rocks on the far western slope of the dome, the region variously termed the Springfield plain, plateau or platform.[1] His second erosional surface consisted of benchlands and terraces carrying old stream gravel, hence his terms "two troughs" and "duplex valleys." Hershey's work in Missouri was limited largely to this Mississippian upland region and, although correct, it fell short of a complete outline of Ozark physiographic evolution. He apparently never realized that the benches and terraces along valleys incised in the Springfield upland were products of the same cycle which almost completely removed Mississippian rocks from the central part of the dome, leaving the lower and much more extensive Salem upland largely on Ordovician rocks. He did not comment on the significant behavior of streams like James River, which flow westward from the (dissected) Salem upland toward the east-facing Burlington or Eureka Springs scarp of the Springfield upland, cross that upland and return to the lower Salem surface.

Later writers[2] who have reported two erosion surfaces on the Ozarks have studied regions where the Salem upland or its equivalent was the higher and older product, where the Springfield surface had been entirely destroyed during the second

[1] Another interpretation of the Springfield upland is that it is a structural plain, commonly referred to as the Burlington plain. But this plain, descending westward at a gentler angle than the underlying Mississippian formations, bevels successively younger rocks until it truncates Pennsylvanian strata. Furthermore, it cuts smoothly across local structures. The Chesapeake fault (Lawrence Co.) has lost 250 stratigraphic feet on its upthrown side and the adjacent Kenoma-Golden City-Miller anticline has suffered a maximum loss of 350 stratigraphic feet from its crest, neither structure being indicated by any break in topographic accordance of hilltops at the Springfield niveau.

[2] Marbut (1907), Buckley (1909), Lee (1913), Dake (1930), Bridge (1930), and Flint (1941).

erosion cycle. Indeed, the writer is convinced that the second base-leveled surface is itself largely gone in the central part of the dome, and that the surface which observers and interpreters of topographic maps have called the Salem upland is mostly the dissected third erosional plain. The best preserved remnants of the second surface are on the cuesta summits and the White-Gasconade divide. If Flint's Ozark peneplain east of the St. Francis Mountains is correctly correlated by the writer, it is the second erosion surface of Ozark physiography, not the earliest.

The extensive erosional reduction in both the Springfield and Ozark cycles justifies use of the term peneplain. The third surface, here called the post-Ozark, consists largely of broad straths, recording very late maturity at the time renewed uplift inaugurated the present cycle. Dake alone has reported an area where the cycle (his second) developed a broad lowland across both headwater valleys and their interfluves.

An element of confusion exists over considerable areas where the resistant Roubidoux formation crops out at the approximate horizon of the post-Ozark surface and determines strongly expressed outer valley bottoms. However, the Roubidoux is known to be beveled along some of these flats and to go below others when traced down along radial stream valleys on both northern and southern slopes of the dome, the post-Ozark surface there being developed on the younger Jefferson City formation.

All students have stated or implied that the pattern of each succeeding uplift departed but little from that of the preceding one. The drainage pattern on the dome has therefore remained essentially the same throughout this history. The remarkably entrenched meanders of modern Ozark streams can have no other explanation (Tarr, 1924) and the lack of windgaps in the cuestas indicates that there have been no notable earlier piracies.

One departure of a later warping pattern from an earlier one is recorded in the valley of the Osage River. In addition to the comment of Tarr that its upper reaches are in old age while its lower portion is in early maturity, it should be noted that its bounding uplands, traced downstream, increase in altitude as much as 100 feet in little more than 50 miles of valley length as the river enters the area of the Ozark dome.

Further evidence of slight departures from earlier uplift patterns is found in varying vertical intervals between the erosion surfaces, probably the most pronounced occurring along the Eureka Springs escarpment where the Ozark peneplain lies 350 feet below the Springfield in southwestern Missouri (Ava quadrangle) but, traced northward, this vertical interval decreases to extinction in about 100 miles, the two surfaces there being indistinguishable.

The vertical intervals separating the different erosion surfaces must be considered in the cave theory herein set forth. These intervals vary because of differential uplift and erosion and perhaps also because of differential solutional lowering in two adjacent surfaces since they were uplifted. Rarely, however, is there a difference of 200 feet in altitude; more commonly the interval is of the order of 100 feet.

The Ozark dome should afford a fair test of the Walther Penck theory (1924) of geomorphic evolution, in which the influence of uplift far transcends that of rock structure. The dome should, by that theory, possess a series of "treppen" consisting of scarps facing outward and retreating parallel to themselves toward the center of the dome, each scarp overlooking an erosional plain lying in front of it. The exact opposite of this dominates Ozark physiography; the scarps being the outcrops of more resistant formations and retreating away from the area of greatest uplift. The Davis concept of physiographic evolution far better fits the topography of the Ozark dome.

Stream gravel, said Fenneman (1936), is the least expectable feature of a peneplain. Acceptance of this statement requires some special explanation for the blanket of chert stream gravel up to 12 feet in maximum thickness unconformably overlying the diaspore and burley-filled sinks well down the northern slope of the dome. The explanation seems to be that uplift of the Ozark peneplain was first felt in restricted central areas and gradually spread radially, rejuvenating the old age streams in the same sequence. Thus this thick and widespread gravel on the Bourbeuse-Gasconade divide on the northern slope and on uplands of White River drainage on the southern slope was an alluviation on yet unelevated flanks of the peneplained dome.

OZARK CAVE-MAKING

As a result of the writer's investigation, 38 Missouri cave maps are in possession of the Missouri Geological Survey, five of them elaborately done with contoured walls and ceilings as well as floors, and the remainder, surveyed by tape or pace and compass, showing ground plan only. Two types of cave patterns appear in these maps, (1) a network of narrow and generally high passageways determined by two or more intersecting sets of joints, and (2) sinuous or subangularly irregular passageways with converging branches of similar character, the whole suggesting a dendritic stream pattern. Most Missouri caves seen are confined vertically within a narrow stratigraphic range, and only a few have more than one "level."

More than half of the caves examined have either perennial- or storm-water streams today and many others record former streams. A few cave streams can be traced to sink hole sources and a very few to engulfment of surface streams. Most of them are dripwater collections or enter the caves through minor bedding plane crevices. To a small extent, these cave streams undercut walls or erode gorges in bedrock floors and thus add locally to a cave's capacity. Far more commonly such streams bury the bedrock floor with detritus or flowstone. Sink hole breaches in cave roof rock are almost all younger than the cave, and debris contributed from them and not carried on by storm water collects in alluvial fans or cones to decrease an earlier capacity. Associated dripstone deposits contribute to the same result. Ground-water changes today are destroying Ozark caves. Surface streams, widening the valleys into which caves open, destroy the lower cave lengths. Waste from valley slopes migrates back into the mouths of streamless caves, in some cases almost obliterating such openings. Failure of cave roof rock is widespread. Most enterable caves of the Ozarks are in their decadence, emphatically are not being made under present conditions.

These destructive changes are normal consequences of the present cycle of erosion. As surface valleys are deepened, the chief underground change in cavernous zones is lowering of the upper limit of constantly saturated rock. If this brings about cave deterioration and destruction, then the caves must pre-date such lowering past their horizon.

This *a priori* argument does not stand alone in defense of the Davis thesis. The writer has published (1942) descriptions of solutional features on cave walls and ceilings and of cave patterns which are essentially impossible by the older idea that streams now flowing on cave floors are the makers of the

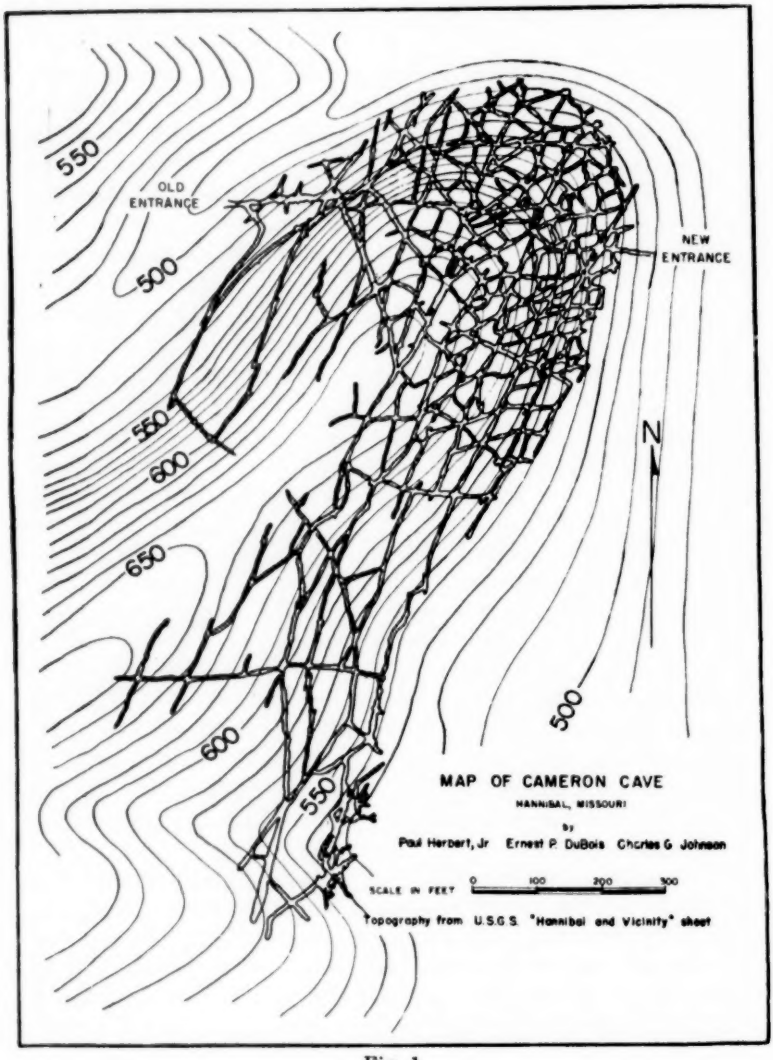

Fig. 1

caves they follow. Instead of briefing the descriptions here, a few typical Missouri caves will be described.

Cameron Cave.—Figure 1 is a carefully surveyed map of Cameron Cave and its overlying topography. Cameron is two miles south of Hannibal, Marion Co., and not far from the well-known Mark Twain Cave, a creek valley separating them. The total length of all of Cameron's passages is four and one fourth miles, the total area above the cave only about 30 acres. The network of high, narrow passages along four sets of joints completely underlies the hill whose slopes transect the cave system in more than 25 places. The system lies low in the hill, is confined to the Louisiana limestone and is roofed by the Hannibal shale. The shale roof is tight and, except at the slope waste blockades where passages and hillsides intersect, there is no seepage water and no secondary limestone in the cave.

It has been suggested that Cameron was made when Mississippi River water, during higher Pleistocene stages, ran through the base of this hill along the joint cracks, its solvent action leaving the amazingly complex network. This idea is rejected because (1) no gradient or pressure adequate to cause such a circulation appears remotely possible, (2) jointed limestone well exposed in quarries and road cuts in the vicinity and within reach of Pleistocene flood waters has essentially no cavern development, and (3) another splendidly developed network cave (Mark Twain) lies in the same stratigraphic position but twice as far back up the tributary valley from the Mississippi as Cameron.

Instead of Cameron and Mark Twain networks taking origin under their respective hills, these hills and the separating creek valley are younger than the caves. The valley development has cut an earlier large network cave in two, Twain and Cameron being only separated remnants. A subwater table circulation, perhaps artesian, followed the joint system down the gentle dip of the formation and dissolved the limestone in places *up* to the overlying shale long before the adjacent Mississippi River trench had been deepened to the cave level. Discharge of such a phreatic flow occurred as a big spring somewhere in the lowland of that earlier topography, a discharge comparable to the pressure flow of many of the Ozark big springs of today.

Vadose water has later used both caves as valley deepening reached their level. The creek itself has found entrance through some of the many hillside intersections and has wandered through limited parts of both caves, cutting non-paired shelves and meander niches in the phreatic spongework walls of the passages it used and bringing in hackberry seeds, snail shells and pebbles from the glacial drift. The entire pattern is utterly impossible for a free-surface gravity cave stream. It required phreatic conditions for its origin.

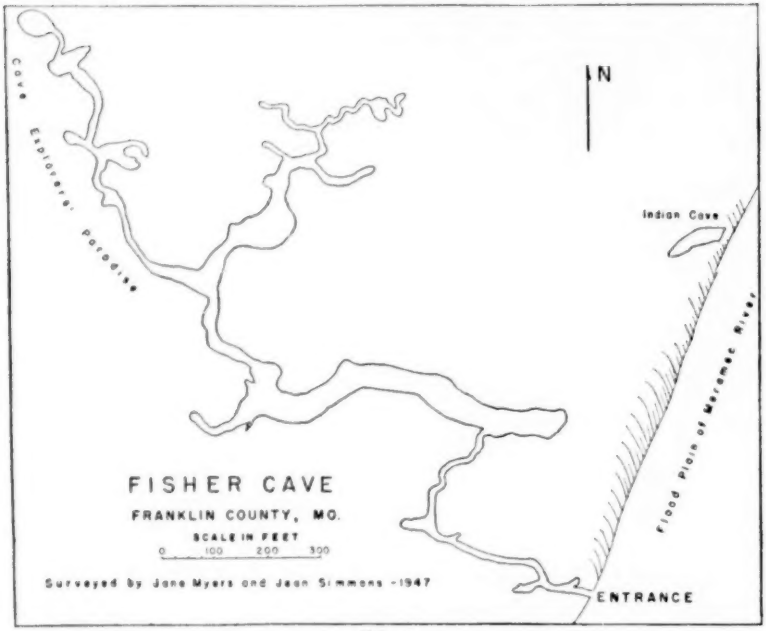

Fig. 2

Fisher Cave.—Figure 2 shows another carefully surveyed Missouri cave, one exemplifying the dendritic type of ground plan. Fisher Cave, in Meramec State Park, opens at flood plain level on the Meramec River valley, its perennial stream supplied by drip and bedding plane seepage and never having flash floods of muddy water such as sink holes feed into caves. Fisher is exceptional, although not unique, in having no wide-mouth hillside opening of the main chamber. Instead, the stream following that capacious linear room abruptly leaves it by a narrow passage 600 feet long, a stoopway for part of

that distance, to reach the outside. The main chamber, headed for the Meramec's cliffed valley side, is completely blockaded by dripstone 100 feet beyond the head of the stoopway. Also exceptional but not unique is a bedrock floor for the 600 foot passage although there is no sign of such beneath the trench cut in the tan-colored clay fill that floors the high-ceilinged main chamber. The surface of the clay deposit is higher than the ceiling of the stream's escape route, the bottom is an unknown depth below. Clearly the small passage is of later development than the main chamber, is probably younger than the tan-colored clay fill. An adequately proportioned valleyward continuation of the main chamber must exist beyond the massive dripstone blockade.

The river cliff is less than 200 feet distant and its base is about 25 feet lower than the clay floor at the blockade. In this cliff, about 700 feet north of the stream's emergence, is Indian Cave, a small opening some 30 feet above the floodplain but containing a fill of unknown depth back of the cliff talus. Indian can be penetrated for 50 feet or so, its course back under the hill heading for the blockaded terminus of Fisher. In all likelihood Indian Cave is a portion of the continuation sought. A straight line connection would mark about 350 feet of occluded main chamber here.

The top of the fill in Indian contains a stream gravel close to the ceiling, the record of a free-surface or vadose stream which functioned just before or just after the fill was completed and certainly before this river cliff was made. Thirty to 40 feet of deepening of Meramec Valley since the gravel was deposited seems indicated. The detrital fill, not the dripstone, is the real blockade. Both caves are older than the fill, therefore are older than the present river valley. By no stretch of the imagination can the present cave stream of Fisher be held responsible for the main chamber with its ceiling of spongework deeply indented with upside down "wash tubs," "bath tubs," "chimneys" and "nail kegs." Its narrow escapeway is the only part of Fisher attributable to its own work.

Back under the adjacent upland, Fisher possesses three converging branches whose gradients rise upstream until at the head of one the cave roof is only 50 feet thick. Here the surface slope descends 60 feet northward in a distance in which the cave floor descends 40 feet southward. The cave stream

thus drains toward and crosses beneath a hill spur and in this distance it flows parallel to the Meramec Valley but in the opposite direction of the river's flow. Fisher Cave is older than the present topography.

The tan-colored clay deposit constituting the essentially horizontal floor extends upstream for 600 feet, to the junction of the first tributary. Farther than this place, the cave bottom rises above the clay floor level. But a clay has filled these upper portions. It is an older deposit, unctuously smooth, strongly red in color, completely devoid of any included chert gravel, sand, dripstone or flowstone. Remnants linger even in ceiling spongework pockets. The "Cave Explorer's Paradise" branch is a most intricately tortuous crawlway, or coalescence of crawlways, where no fat man will ever verify the statement that only partial clearing of a former complete red clay fill has occurred. This is the peneplain clay referred to in the abstract, once filling the highest and most intricately irregular cave openings of phreatic origin and now being removed by vadose water in the present erosion cycle. The tan-colored clay was deposited only after most of the red clay had disappeared and after considerable stalagmitic growth had occurred. The flat floor is its upper limit.

The cave was completely filled with ground water during its growth and it remained completely filled when solution was succeeded by deposition of the red clay. No Meramec Valley like the present one could have existed during either of these episodes. The solutional episode is assigned to the maturity of the Ozark cycle of erosion when hydrostatic head under uplands could provide for the subwater table circulation that made the cave. The episode of red clay filling marks the attainment of the Ozark peneplain, upland reduction having destroyed the hydrostatic head by which that deep movement of ground water toward the ancestral Meramec made most of Fisher Cave.

The Meramec State Park quadrangle map shows that the hill spurs close to the river valley have flattish tops between 800 and 840 feet A.T. Farther back from the river is a dissected plain, the summit country, approximately a hundred feet higher. If these surfaces are remnants of the Ozark peneplain and the post-Ozark strath and if Fisher is of phreatic origin, it follows that the cave, ranging in altitude between 600 and 690 ft. A.T., was made as much as 200 feet below the post-Ozark

strath, or 300 feet below the Ozark peneplain. The ancestral Meramec theoretically received the cave's phreatic discharge by way of a big spring operating under hydrostatic pressure.

Tunnel Cave.—Appropriately named, this cave perforates Bear Ridge, a part of a hill spur occupying one of the great incised meanders of Gasconade River in Pulaski County. It receives storm water from a minor ravine, once tributary to the Gasconade and, although one of the few Ozark caves exemplifying Malott's invasion theory, it retains a perfectly adequate record of its phreatic origin.

Subterranean piracy of the ravine drainage occurred at the east (intake) end of the tunnel when the cave roof directly beneath the ravine failed and formed a ponor (collapse sinkhole) 75 feet deep. The stream, plunging into this and following the cave, reached Gasconade River floodplain on the other side of Bear Ridge, 1000 feet or so distant. In so doing, it abandoned more than a mile of its lower course and developed a sharply intrenched little rockwalled gorge in the ravine upstream from the ponor.

Ceilings and upper walls of Tunnel Cave carry excellent spongework and some can still be recognized on lower walls although greatly abraded by gravel carried through in flood time. Few caves of the writer's experience show so well a young, canyon-like, vadose, lower part of the cross section cut into the bottom of a wider, older cave retaining phreatic traits.

Tunnel antedates the valley-making and ravine-making of the present cycle, therefore is older than the ridge it perforates. Only after the river had deepened to the cave level could vadose conditions obtain. Only after the ponor collapse occurred could a surface stream enter.

Remnants of the Ozark peneplain on the summit of the cuesta a few miles north of Tunnel Cave lie close to 1200 feet A.T. Intermediate flattish summits closer to the river and presumed to be post-Ozark strath remnants are approximately 1000 ft. A.T. The cave mouth is about 800 ft. If the phreatic history of Tunnel dates back to the Ozark cycle, cave-making occurred here 400 feet below the surface of the finally attained peneplain.

Railroad Cave.—Hardly more than two miles from Tunnel is a now streamless cave with a quite different record of surface stream invasion of an original phreatic route. This is Railroad

Cave, difficult to find in the ravined and wooded country nearer the river in the same hill spur but the most outstanding cave in the writer's experience for showing superimposition of vadose stream meanders on magnificent spongework walls of phreatic origin.

Railroad Cave is not far from horizontal and has only a gopher hole entrance at each end, dug through hillside waste. A short distance inside, beyond the in-sloping talus of surface debris, Railroad is a capacious single corridor more than 1600 feet long. Although without branch chambers, it possesses alcoves along its sides, each alcove an upright half-cone cut into the spongework walls, each possessing an "island" of rock also rudely half-cone in shape and conforming to the alcove ceiling slope.

These islands are slip-off slopes and the alcoves are half-cone meander niches (see Bretz, 1942, pp. 679-682) left by a free-surface stream flowing on, and gradually lowering, a clay fill. In Railroad such niches or alcoves incise the older chamber wall in no less than 29 places, generally alternating with each other on the two bounding walls. Many of these meanders began their incision and downward enlargement close to the ceiling, others took origin at lower levels as the vadose stream lowered the clay floor. Stream pebbles still lie on some of these meander floors. The fill on which the vadose stream flowed is the red peneplain clay, much of which still remains, even in ceiling spongework cavities.

Railroad's termini are well up on the hill slopes. The cave perforates a ridge on which there is a minor ravine groove. This fails to reach the cave but its water leaks down into the cave and has made the only noteworthy dripstone deposits Railroad possesses. Solution also has occurred at this leakage and produced another vadose superimposition on the phreatic cave, an incipient dome-pit (see Bretz, 1942, pp. 682-685).

The vadose stream which added the half-cone meander niche alcoves came from one of the ravines outlining the ridge and came early in the ravine-making. The record is of a relatively short-lived subterranean piracy, the stream later returning to its outside course. All this vadose alteration belongs to the present cycle but the spongework cave had to exist previously. The red clay fill, although later than the phreatic cave, also

had to be made before the bounding ravines were incised and the vadose stream entered the cave.

If Railroad Cave's phreatic history is to be correlated with the Ozark cycle, the cave was originally developed 300 feet below the peneplain but, by the argument of this paper, it was made before the peneplain stage had been attained, therefore at a greater depth.

OZARK BIG SPRINGS

By Meinzer's definition (1927), a first magnitude spring has an average discharge of 100 second feet (64 million gallons a day). Beckman and Hinchey (1944) found that Missouri possesses 12 such springs, and half a dozen more almost as large, all in the region of the Ozark dome. There can be no question of the existence of an integrated and capacious underground drainage system to supply these springs. The steady flow and lack of turbidity except after heavy rains indicate discharge from considerable subwater table reservoirs. Emphatically, they are not resurgences of lost rivers and sinking creeks such as Malott found in Indiana.

These large springs emerge at the bottoms of deep Ozark valleys, some of them even in the bottom of the river channel, and their present mouthings cannot date very far back in the present erosion cycle. Either their underground drainage systems have developed *de novo* since the river valleys reached their present depths, or they are supplied by phreatic cave systems older than present topography and only recently incised by valley deepening.

The writer knows of four large Missouri springs discharging from enterable caves. The mouth of one of the Greer springs, Oregon County, can be entered by boat or by wading and Roaring River spring, Barry County, only by boat but in a short distance in each the ceiling descends to the water surface of even the lowest discharge. Welch and Fishing Cave springs, Shannon County, rise in the middle of cave floors and soundings in them are surprisingly deep and very instructive as to cave theory.

Some of the big springs emerge under pressure. Big Spring, Carter County, the largest of all Ozark springs, is the outstanding example.

Big Spring.—The up-rushing discharge of this great spring throws spray above the top of the markedly turbulent water dome where an average of 252,000,000 gallons per day escape to nearby Current River. The spring has never been sounded, perhaps never can be because of the spectacular vigor of upwelling. Such a discharge is judged to be from below a water table level in adjoining higher country of adequate area, and to be using a cave system akin to that of Fisher Cave but not yet drained by sufficient deepening of Current River valley as has been done for Fisher by Meramec River. The Big Spring "Cave" is still full of water to the very ceiling, at least in its lower stretches. Its phreatic history has not yet entirely ended; it has not yet become vadose throughout.

Wastes from an iron furnace some miles distant have contaminated the spring in the past (Bridge, 1930). To do this, subterranean drainage had to *cross under* Pike Creek valley.

Roaring River Spring.—The discharge of 6½ to nearly 70 million gallons per day marks Roaring River Spring, Barry County, as a large spring, if not of the first magnitude. Situated near the convergence of two canyon-like valleys nearly 500 feet deep, the spring is the head of Roaring River during dry weather. It is on the east side of the smaller of these valleys and has no surface valley of its own. Conditioned by other nearby deep, steep-sided valleys, the spring cannot receive more than a small share of local ground water, from about two square miles on that side. The total surface drainage area that might supply local ground water from the *opposite* side of the valley is 33 square miles, and here equally deep and steep-sided valleys must take the lion's share of the precipitation. Roaring River Spring must secure its supply from outside the surface drainage area.

The region is a part of the deeply dissected country constituting the southeast-facing Eureka Springs escarpment. Most of the surface drainage on the back slope of the cuesta flows northward, away from Roaring River. Large areas on this back slope are little altered remnants of the Springfield peneplain. Depths locally of 150 feet of cherty waste are known on it. A large shallow doline on this plateau, five miles distant from the spring, 350 feet higher and containing a lake of several acres, abruptly lost its standing water in 1939 or 1940 (Beckman and Hinchey, 1944), and Roaring River Spring equally

abruptly became "very muddy for several hours, causing a near-disaster in the rearing pools for baby trout" at the spring mouth.

Thus we know that the spring's supply comes from the cuesta's back slope, moves in nearly the opposite direction of the surface drainage and goes *under* Roaring River Hollow in order to discharge on the east side of the valley. Only the existence of a pre-canyoning phreatic route can fit all these facts. That route is still largely phreatic. And it lies nearly 500 feet below the peneplain remnants on the cuesta brink.

Welch Cave and Spring.—This, the fifth largest spring in the Missouri Ozarks, discharges into Current River Valley in Shannon County, on the south slope of the Ozark dome. Facts regarding it must be taken largely from other sources. The cave can be entered only by boat and no boat or raft was obtainable at the time of the writer's visit. The spring, according to Beckman and Hinchey (1944), discharges a minimum of 50 million gallons per day, a maximum of more than 200 million. They agree with Doll's idea (1939) that it is supplied by ground water from the Meramec drainage on the north slope of the dome. This drainage must cross under the Meramec-Current divide, a situation even more pronounced than that of Roaring River Spring.

Welch Spring is a great upwelling in the middle of the cave floor. Farrar (unpublished notes, about 1940) said that the "lake" made by the spring, a short distance inside the cave mouth, is 200 feet across and 75 feet deep. Beyond the lake is an "auditorium" 120 by 120 and 75 feet high, and "many side tunnels and crevices," but no stream. With the high, slot-like passage carrying the spring water to the Current River bluff, this all strongly suggests a phreatic cave system.

Although Current River's muddy flood water backs up in the cave, the 75-foot hole remains unfilled, and it seems probable that the spring is fed from a level 65 feet below river bottom. The existence of this hole is the most significant fact about Welch Cave and Spring.

The enterable part of Welch Cave is in the vadose zone but its stream rises from below the water table, is phreatic water from deeper, still submerged portions of the system. As with Big and Roaring River springs, the underground drainage system supplying Welch Spring cannot have been developed

since Current River valley reached its present depth. The Ozark peneplain remnants on the Meramec-Current divide, a few miles to the north, are 500 feet or so above that valley bottom. If the cave system is of phreatic origin, it should date back to maturity of the Ozark cycle or at least to that of the post-Ozark partial cycle, and vigorous ground water circulation 400 to 500 or more feet below the surface must be envisaged.

Fishing Spring Cave.—Similar to Welch and in the same county and river valley, all of Fishing Spring Cave's air-filled portion can be seen (but not reached) without use of a boat. The cave opens at the foot of a sheer cliff at ordinary river levels. During low water a strong current three or four feet deep flows out of the cave over a rock floor. In high river stages, the cave is entirely submerged.

It is a surprisingly short cave above water level, ending rather abruptly less than 150 feet back of the cliff face. Except for two rock shelves, the entire cave floor is under water. The discharge is from 16 to 47 million gallons daily.

This water rises at the back end of the cave, making a low mound off which the boat constantly tended to slide while soundings were being made. Two lateral soundings were 73 and 74 feet and a central one was 77 feet below low water stage of the spring. A 32 candle power spotlight showed vertical walls far down on all but one side where there was a submerged overhang, the extent of which was indeterminable.

Mud from Current River floods covers the two rock shelves but the deep hole remains unfilled, due presumably to rise of spring water from the very bottom, more than 60 feet below Current River floodplain.

The river apparently has never had a deep stage, now alluvium-filled to the floodplain level, and this vertical rise records a joint-controlled passage demanded by phreatic water under the Salem (Ozark and post-Ozark) upland to the north. The cave passages submerged in Fishing and Welch are now, and always have been, in the phreatic zone. If this is true, then clearly a phreatic circulation may demand upward components near its debouchure. Certainly the underground conduits leading to Fishing Cave Spring antedate the attainment of present depths of Current River valley.

The big springs above briefly described are interpreted as still functioning portions of the deep phreatic drainage system

developed during a former erosion cycle that eventually peneplained a structural dome largely of calcareous rock. The enterable caves of the dome are portions of that system lying high enough in the rock to have become drained during the later stream rejuvenation. Most of these air-filled caves possess remnants of a red clay fill deposited subsequent to the phreatic cave-making but now being removed by vadose water using the cave route. The clay fill is interpreted as a record of the peneplain stage of the earlier erosion cycle. All members of the phreatic system should once have received such a fill. In the present cycle's rejuvenation of surface streams, many properly situated subterranean conduits have also undergone "rejuvenation." The big springs of today are members of that group. From the thorough flushing they now receive, all red clay they once contained must be gone. Such subterranean rejuvenation has probably been responsible for complete or nearly complete removal of the red clay fill in some Ozark caves now air-filled and possessing only a trifling vadose stream on their floors.

MECHANICS AND DATING OF THE DEEP PHREATIC CIRCULATION

Widely accepted interpretations, especially if entrenched in textbooks, often become authoritative dogma, against which printed descriptions supporting challenges of the established view may make little impression. Davis was aware of this inertia when he wrote on "the value of outrageous geological hypotheses" (1926), and when he asked that caves be re-examined to check on the validity of the theory of vadose origin. Proponents of that theory may grant all the criteria he and later students have argued to be the results of a complete water-filling, but ask that they be explained as flood-time products under present conditions. The existence of the unctuous red clay fill may be granted but held to be due to local vadose pondings. Granted the absence of a karst topography; cannot a karst have existed earlier, cannot the caves be vadose from an earlier cycle, losing their karst during a later peneplanation? Don't all dendritic caves lead to present valleys as vadose caves must do?

There are other challenges of the Davis concept possible for defenders of the orthodox view. One that Davis rightly considered fundamental but could not answer with definite facts

was demand for an adequate mechanics for the deep circulation he envisaged. Joint and stratification partings exist in the saturated zone but a circulation along them, eventually producing an integrated subwater table cave system, would require that the water escape *upward* to enter surface stream valleys.

Davis had recourse to a theoretical picture drawn by King (1899) for ground water movement in homogeneous permeable rock under a topography of adequate relief. Because the water table stands higher under hilltops than under valley bottoms and, as Davis added, because most sedimentary formations are cut by intersecting joints and bedding planes, the weight of the skeletonized water column under a hill should produce a deep lateral movement toward an adjacent valley. Curved courses of such flow, (see fig. 3) zigzagged to fit the pattern of the partings, would go deeper than the valley bottoms, then rise as the escape route was approached. With well-developed integration of the solutionally enlarged, favored courses, large springs would exist in those valley floors, their water rising along joints. This idea of deep flow under hydrostatic pressure, without using the added idea of solutional enlargement, has found favor with Slichter (1902), Tolman (1937) and Hubbert (1940) in their ground water studies. The writer had adopted it long before he learned of Welch and Fishing Cave springs, rising vertically from depths of some 70 feet below valley bottom.

There are numerous recorded instances of drillers' bits abruptly dropping into cavities below the region's water table, this despite Gardner's skepticism (1935). Moneymaker (1941) has recorded such solution cavities in bedrock at dam sites along Tennessee River, encountered below rock bottom level of the river. Well drillers have found similar cavities in several places in the Ozarks, some of them water-bearing (Beckman and Hinchey, 1944, p. 25), some yielding a red muddy water (E. L. Clark, personal communication).

Ozark springs and caves fit this concept of flow under hydrostatic pressure. Scarcely more than half a dozen of the 138 Missouri caves studied lack all evidence of phreatic origin and development. Recording a marked integration of ground water flow in the saturated zone, and in many cases oriented at marked variance to control by existing hills and valleys, they must date back to an earlier, higher water table which has been destroyed in making the present rugged Ozark topography.

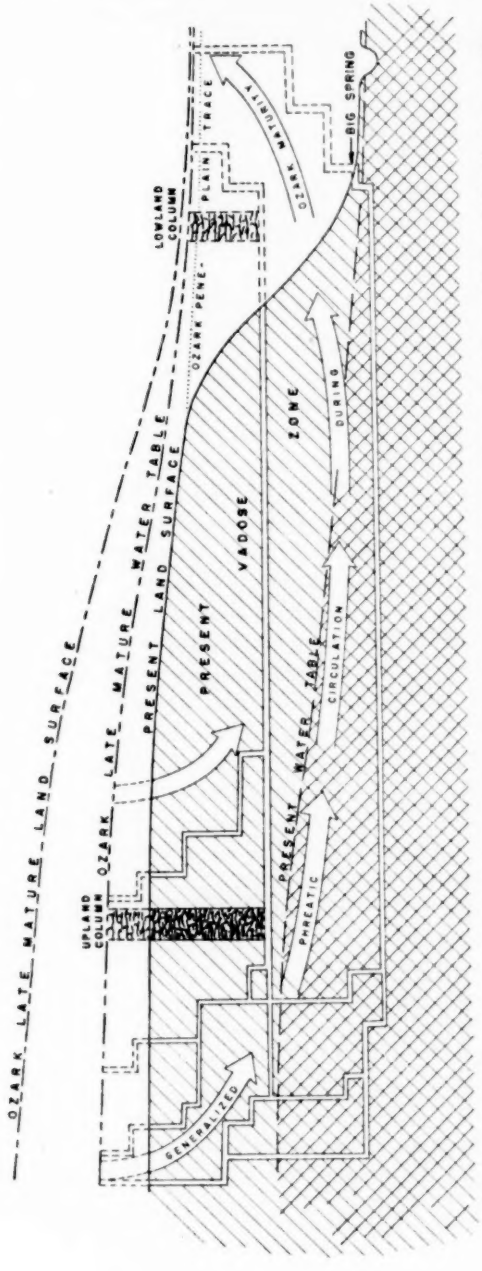

Fig. 3. Diagrammatic depiction of sequential changes in ground water circulation during evolution of Ozark topography.

These caves have suffered complete destruction of their lower stretches because of this later valley deepening. The missing parts must perforce be the theoretical rising water routes, leading to the theoretical large springs of an earlier cycle. The last cave examined during the field study happened to be Fishing Cave with its upwelling big spring. The writer's reaction to the soundings obtained can be readily imagined. At Fishing, the downcutting of Current River in the present cycle has been slightly offside a part of the rising route for that early hydrostatic flow, and in addition has thus far failed to deepen to the main horizontal component of that flow. The cave of the earlier cycle therefore is still full of water flowing vigorously under completely saturated conditions. Welch appears to be a similar case. Elsewhere, so far as now known, the present cycle of erosion has destroyed the upward components of the zigzag curved courses the theory requires. Some of the mounded or "boiling" big spring discharges, when sounded, may add to this list.

Only a region of marked relief can provide the hydrostatic head for phreatic cave-making. Today's Ozark topography has the necessary relief and possesses phreatic springs. Why should not all the abandoned caves of subwater table origin be ascribed to earlier stages of the *present* cycle? There are two answers to this challenge. One is that some caves are so high in the interfluves that peneplained flats are just above them. No adequate relief existed during early youth of the present cycle for providing the phreatic circulation they record. Nor can they be considered as products of ground water circulation under the base-leveled conditions which left the peneplain. That circulation was almost nil.

It is the red clay which records the peneplain ground water. It is this clay which provides the second answer to the challenge. There is no other time in Ozark geomorphic history for this almost universally recorded experience of the phreatic caves. Certainly the present cycle provides no suggestion of such a time.

The last item to be considered in this study is the chronological relation of the Ozark phreatic caves to the three earlier erosion cycles. Which caves belong to which cycle? It seems probable that, with each succeeding uplift and consequent reduction of the land surface, caves would be made at succes-

sively deeper levels in the calcareous rocks. But no criterion for identification of three possible generations of caves appears to exist. Depth below identifiable erosion surfaces might seem a possibility until it is realized that the vertical range between the Springfield and the Ozark, or between the Ozark and the post-Ozark base-leveled surfaces everywhere is less than the depths at which some phreatic caves were made. Furthermore, the depth at which circulation was adequate depended in part on where ground water found easiest lateral movement toward valleyways. Hydrostatic head also varied, from place to place, in terms of the overlying topography. Thus, if there were successive generations of caves, they undoubtedly overlapped. Only those caves shallowly located under remnants of the Springfield peneplain can be definitely dated.

Caves developed but little below the water table of the earliest cycle would presumably receive a red clay fill during the peneplain stage and then pass into the vadose zone by maturity of the second cycle and so remain, if they have survived, high in the Springfield uplands. If they were made far below that water table, they could have remained phreatic during the next cycle, though perhaps losing their clay fill and undergoing further phreatic solution if they became, like today's large springs, routes of vigorous subwater table flow. Conceivably they could have received another red clay filling during old age of the second cycle, and have passed into the vadose zone during the post-Ozark or even the present cycle.

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LIFE AND DEATH ASSEMBLAGES AMONG FOSSILS*

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ABSTRACT. Discrimination of fossil bivalve life assemblages and death assemblages comparable to Recent biocoenoses and thanatocoenoses is difficult. Size-frequency distributions have been analyzed in terms of the biocoenosis of the following: wave action, solution and abrasion, lower survivorship curve, in addition to the biological and physical factors tending to modify the initial distribution. The ratio of articulated shells to disarticulated shells is considered to be a function of time and to be related to the ease with which the shell may be disarticulated by biological and physical agencies. The departure from unity of the ratio of matching opposite valves is thought to be caused by the disruptive effects on the biocoenosis of the following: wave action, solution and abrasion, lower limit of observation, selective action of predators, scavengers and boring or triturating organisms. Comparison of the population densities observed among living forms may aid in differentiating between life and death assemblages among fossils. Application of some of these criteria to assemblages obtained from the Lower Devonian of northern Maine suggests the presence of five death assemblages plus one life assemblage.

INTRODUCTION

A paleoecologic study should attempt, as far as possible, to reconstruct the total environment for the particular time and place in question, including the physical and biological factors that affect the life and relationships of organisms.

Research in ecology has shown that detailed studies are necessary for understanding the relationships between organisms and their environment. In such studies the term "biocoenosis" has been used. A biocoenosis, meaning community of life, comprises the living organisms situated within a particular area. Much of the work of ecologists interested in marine benthonic organisms has centered upon the composition and functioning of the biocoenosis.

The paleontologist is handicapped in ecologic studies by the necessity for distinguishing ancient biotic assemblages comparable to the biocoenosis from what Wasmund (1926) has termed the "thanatocoenosis." Strictly speaking, the term "thanatocoenosis" implies a community of death; as used by Wasmund, however, it has come to mean the aggregated remains

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of organisms that in many cases never constituted a biocoenosis, but were brought together by physical agencies, such as wave and current action, which operated after they died. An excellent example of a thanatocoenosis is the miscellaneous assortment of organic debris gathered on most marine beaches. The fact, however, that fossils are commonly not preserved in their proper habitat need not deter the paleontologist from seeking to differentiate between assemblages comparable to the biocoenosis and the thanatocoenosis of the neontologist.

It is here suggested that the terms "life assemblage" and "death assemblage," respectively, be employed as the paleontological equivalents of biocoenosis and thanatocoenosis.

Clearly, the first task facing the student of paleoecology is to distinguish between associations comprising a life assemblage and those making up a death assemblage; this requires detailed observation and comparison of results with those obtained in studies of Recent ecology.

One of the best ways to determine the nature of a community is to subject certain of its properties to statistical analysis. Statistical analysis is an extended means of observation, the results of which enable the worker more easily to comprehend the type of community with which he is dealing. This paper is an attempt to deal statistically with assemblages of marine bivalves, referring especially to fossil brachiopods.

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CRITERIA FOR DISCRIMINATING BETWEEN A LIFE ASSEMBLAGE AND A DEATH ASSEMBLAGE

To interpret the results of this statistical analysis, one must first consider what conditions will be of importance in producing any given assemblage of hard parts of marine bivalves such as brachiopods.

Size-frequency distributions.—The simplest example of a size-frequency distribution is that of the assemblage of hard parts

left behind by a single set[1] at a given locality. The two primary circumstances governing the size-frequency distribution of any set are the rate of reproduction and the death rate. In addition, the vagaries of the transporting current that brought the planktonic larvae to the locality where they finally became members of the benthos must be considered. The following discussion does not take into consideration the possibility of sexual dimorphism among dioecious species, such a phenomenon probably being too rare to be significant in most cases.

Many marine invertebrates, especially among the pelecypods, have reproductive and initial death rates that are high compared with those of mammals (Deevey, 1947). It is probable that many brachiopods and pelecypods have a concave or steeply negative sloping survivorship curve, such as that inferred for oysters by Deevey (1947, p. 312). Of importance to the paleontologist is the point on the survivorship curve at which the organism first develops hard parts capable of being preserved as fossils. Percival (1944) presented graphs of size-frequency distributions for the living brachiopod *Terebratella inconspicua* that indicate a survivorship curve that is concave. Under normal conditions, therefore, one may expect a size-frequency distribution curve for individuals past the larval stage that is concave, representing a large number of immature individuals, a moderate number of adults, and a relatively minor number of gerontic individuals, or a steeply sloping negative curve that approaches the asymptotic only for the gerontic individuals as is indicated by the right-hand portion of figure 1. If the larvae are included in the study one will have to consider the left-hand portion of figure 1 that is asymptotic to the vertical coordinate. The size-frequency distribution of a life assemblage that has experienced optimum conditions of preservation and no physical disturbance should approximate its mortality curve.

Most deposits contain more than one set, so one must consider the additive effects of sets with different population densities upon the expected size-frequency distributions of the final assemblage. Three types of increment are to be expected: static,

[1] The term "set" as used by the marine biologist refers to the larvae, of one species, that settle down on a particular portion of the bottom after they have passed through a planktonic stage. The larvae spawned at any one time and place may be the source of the set at a number of localities, their distribution being dependent upon the transporting currents. Set is not synonymous with rate of reproduction.

cyclic, and irregular. For all three types it is assumed that the mortality rate will be of the same order of magnitude, especially among the immature individuals, excluding the larvae from consideration. Assuming the mortality curve to remain

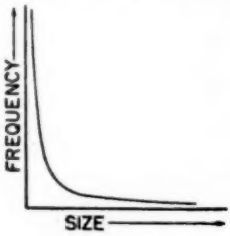

Fig. 1.
Normal survivorship curve.

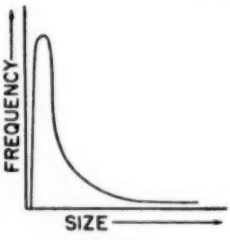

Fig. 2.
Mortality curves for two sets with differing initial population densities and unvarying mortality rate.

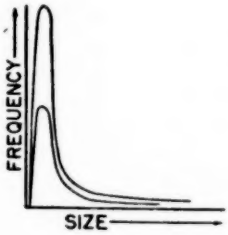

Fig. 3.
Mortality curves for sets with differing mortality rates.

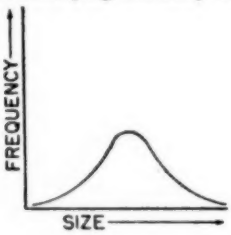

Fig. 4.
Bell-shaped normal distribution curve.

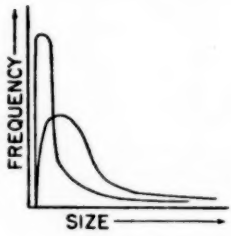

Fig. 5.
Predicted size-frequency distribution for a life assemblage.

about constant, the size-frequency distribution for all three types will be similar to that for the single set, except that the size of the modal class will be exaggerated by the cumulative effect of the additional sets (fig. 2).

The cumulative effect of additional sets with differing mortality rates introduces a slight modification of the size-frequency distribution. This modification will result in right-skewed, concave curves having peaks with variable curvature (fig. 3) as opposed to those showing an unvarying mortality rate. It is conceivable that under certain rare conditions the mortality curve may approach the bell-shaped condition rather than being strongly skewed to the right, with a corresponding effect upon the size-frequency distribution. Such a bell-shaped curve would imply that most of the members of a set had reached at least moderate size and age, an occurrence that seems to be almost unknown among marine bivalves under natural conditions, in which the mean life span is short.

Many secondary conditions, however, tend to alter the expected size-frequency distribution of hard parts. These conditions may be divided into two main groups: (1) non-selective conditions, such as the availability of food, temperature, and disease, all of which will ordinarily affect the size-frequency distribution of all the living forms in an area, and (2) selective conditions that will affect one size group more than another. These latter conditions may be further subdivided into the following categories: physical conditions, such as wave action, current action, solution, abrasion, biological conditions, such as the action of predators (Turner, *et al.*, 1948, pp. 43-45), scavengers, triturating organisms responsible for the destruction of hard parts, and hermit crabs; and experimental limitations such as the practical limits of observation. An attempt will now be made to discuss each of these selective conditions and to assess its effect upon size-frequency curves for the life assemblage and recognize its role in shaping the curves for death assemblages.

Wave action and current action, especially in the littoral zone, are potent agents in the destruction of the biocoenosis, both during the normal life cycles of the various organisms and after their death. Storm waves tend to disrupt the benthos either by burying it or by tearing its elements loose and scattering them. The effect of burial will be, presumably, to preserve a portion of the biocoenosis as a life assemblage. However, the scattering of the benthos will subject its elements to the same sorting processes that affect the accompanying sediments. Wave and current action will continually tend to remove the

hard parts of dead individuals whose competent velocities are less than the competent velocity of the associated sediment (Menard and Boucot, 1951, pp. 146-148) and to redeposit them elsewhere at the dictate of local currents. This removal of hard parts will greatly alter the expected size-frequency distribution, because the greatest number of deaths will occur among the very young members of a set owing to their greater primary abundance. The result may then be a biocoenosis in which the size-frequency distribution curve may approximate a bell-shaped normal distribution[2] and a thanatocoenosis deposited elsewhere which will have the same type of distribution curve owing to sorting (fig. 4).

The effect of solution upon marine bivalves is not well understood. However, smaller hard parts having the same form as larger shells, will suffer disproportionately owing to their greater surface exposed per volume of shell material. This effect will tend to reduce the degree of right-skewness and slope of the curve expected in the size-frequency distribution of a life assemblage.

The effects of abrasion are hard to predict in the absence of adequate experimental data, but it is probable that they would tend to alter the expected size-frequency distribution.

An important factor that must be considered in paleontology is the lower limit of shell size that the methods of sampling and observation allow. In other words, just how small a specimen can be detected in the material? How finely can the sample be broken down in order not to miss any specimens, especially in a heavily indurated matrix? The answers to these queries probably vary for each worker and for each type of material, but some estimate of their effect needs to be taken into account.

The lower limit of shell size preserved may be controlled by

[2] Such an assemblage was observed by Mr. Henry Stetson, Mr. Sherwood Tuttle, and the author, during the spring of 1951, in a salt-water pond at Cohasset, Mass. This pond, connected with the sea by a tidal sluiceway, contains large colonies of the mussel *Mytilus edulis*. Certain of the mussel beds are swept by currents competent to remove all the dead shells, and are underlain by relatively clean-washed sand and gravel. Others are apparently washed by weaker currents, capable only of removing the shells of small individuals. In these latter beds the dead shells of the larger mussels (approximately 3-5 cm in length) are buried below the thick mat of the living in a black, stinking mud. These deposits would give rise to a life assemblage characterized by a bell-shaped size-frequency distribution curve.

the grain-size of the enclosing sediment or the type of alteration (silicification, pyritization).

Two other conditions whose relative importance is not yet understood are the selective action of predators and scavengers in the abstraction of specimens of various sizes from the fauna, as well as the size preferences of boring and triturating organisms.

These modifying conditions indicate that a life assemblage may have a size-frequency distribution that is strongly right-skewed, negatively sloping, showing the predominance of small individuals (fig. 5). Their size-frequency distribution, therefore, may produce a strongly right-skewed or negatively sloping curve rather than the bell-shaped normal distribution curve common for particles of sediment and other materials that have undergone extensive sorting. However, death assemblages whose mode occurs among the small shells, i.e., transported assemblages comprised mainly of small shells, may strongly simulate or even be identical with the above-mentioned right-skewed size-frequency distributions.

The ratio of articulated shells to disarticulated shells.—Marine bivalves tend to become unhinged after death, some groups becoming unhinged more readily than others. Pelecypods equipped with a ligament or a resilium are prone to become unhinged after death, whereas many articulate brachiopods resist this tendency.

The biological and physical circumstances related to this tendency to become unhinged must be considered in interpreting the ecologic significance of the manner of preservation. Governing this event are such biological characteristics as the nature of the articulation between the valves and the rapidity with which the soft parts decompose. It is inferred that size and other biological characteristics are factors within a given species, although these points are not considered in the present paper.

Organisms that live deeply buried are more apt to be undisturbed after death than those that live on the sea floor. However, the presence of growth stages at different levels below the sea floor would tend to disturb the expected size-frequency distribution.

The chief physical conditions that affect unhinging are

the velocity and the duration of current to which the shells are subjected after death (tending to unhinge the shells) and the rapidity of burial immediately after death (tending to prevent the valves from becoming unhinged).

It would be desirable to evaluate these biological and physical conditions among living bivalve faunas. In the absence of such information, an empirical means of evaluating the tendency of bivalves to become unhinged is found by counting the number of articulated shells and disarticulated shells or their impressions for each species in a particular assemblage.

Species that are easily disarticulated will be more delicate indicators of the amount of wave and current action that an assemblage has undergone than species that become disarticulated with difficulty. The occurrence of a single assemblage with a large percentage of articulated specimens of a species that is easily disarticulated would indicate a probable life assemblage. For example, in Devonian sandstones of west-central Maine the pelecypod *Nuculoidea* is usually found only as single valves. This is not remarkable because the resilium would tend to spread the valves apart after death. At one locality, however, a large number of valves are still hinged together, indicating the presence of a life assemblage that was buried before the resilium opened the valves.

A species which is disarticulated with difficulty will be a better indicator of the time during which disarticulating processes have been operative. The percentage of single valves is a function of the time elapsed.

It may be possible to determine the direction of shell movement at a given stratigraphic level by contouring values obtained for the ratio of whole shells and single shells.

The ratio of opposite valves.—An undisturbed deposit of shells containing unhinged shells should yield approximately equal numbers of opposing valves. The opposing valves should have dimensions that permit them to be fitted together. Sorting, on the other hand, will tend to segregate disarticulated valves according to their hydrodynamic behavior.

Field observations show that the opposing valves of different bivalve species may have widely variant hydrodynamic properties. Occasionally specimens of only one valve are found at a given locality; the other valve, because of differing hydrodynamic behavior, has been deposited elsewhere. A sorted de-

posit containing disarticulated bivalves will not, as a rule, yield equal numbers of matched opposing valves.

An empiric measure of these relationships can be obtained by counting the opposite valves in any particular shell deposit. Species whose opposite valves tend to be easily separated from each other will be more delicate indicators of the amount of sorting a deposit has undergone. A deposit containing nearly equal numbers of matching valves of a species whose valves ordinarily tend to become easily separated suggests a life assemblage.

By arranging in order of magnitude the opposite valve and articulated- to disarticulated-shell ratios of any one species taken from different fossil deposits, one may observe whether the same ratios for the accompanying species in each deposit show the same order or whether they vary erratically. If the ratios of the accompanying species in each deposit vary in the same manner as the ratios of the first species, the contents of each bed may have been derived from one source area. If the ratios of the accompanying species do not vary in the same direction as those of the first species at one or more of the deposits, diverse sources may be inferred.

Population densities.—The differentiation of a life assemblage from a death assemblage is aided by study of population densities on individual bedding planes. The population density of the fossils contained in a column of rock is a function of the rate of sedimentation, and consequently is of little value in discriminating between life and death assemblages. Normal population densities vary with the size of the growth stages of each species. For example, Percival (1944) in a biocoenosis of the brachiopod *Terebratella inconspicua* found about 400 individuals up to 3.5 mm long per 2.3 square decimeters and 30 to 40 individuals 10.5 to 13.5 mm per 2.3 square decimeters. Therefore a thanatocoenosis or death assemblage might be indicated by the occurrence of, for example, 160 individuals of this species with lengths of 10.5 to 13.5 mm per 2.3 square decimeters. This interpretation would be strengthened if the valves of this and other growth stages were disarticulated (assuming that this species is not easily disarticulated), if there were widely differing numbers of matching valves, and also if the size-frequency distribution were bell-shaped.

Another approach to this problem might be made by a

study of the hydrodynamic behavior of the remains of benthonic organisms, in order to ascertain their competent velocities. Little work has been done in this direction (Menard and Boucot, 1951).

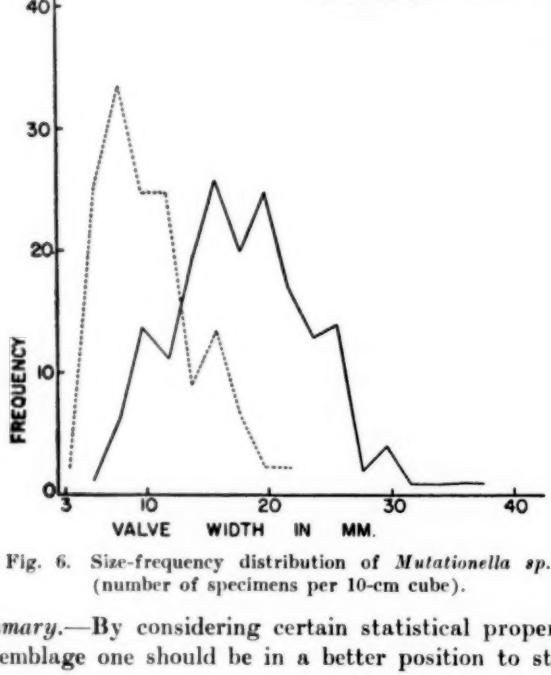

Fig. 6. Size-frequency distribution of *Mutationella sp.* (number of specimens per 10-cm cube).

Summary.—By considering certain statistical properties of an assemblage one should be in a better position to state the

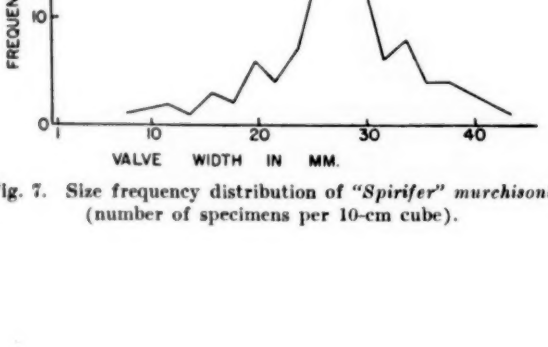

Fig. 7. Size frequency distribution of *"Spirifer" murchisoni* (number of specimens per 10-cm cube).

probability that the population from which the sample was drawn once formed part of a biocoenosis.

By considering also other pertinent information, such as that obtained from a study of the petrography and composition of the entombing sediments, one should be able to add further evidence. Of particular interest is the sulfur content. Sediments that run high in sulfur may have been deposited under reducing conditions, with consequent low oxygen content, and probable slight current activity, and therefore any benthonic animals

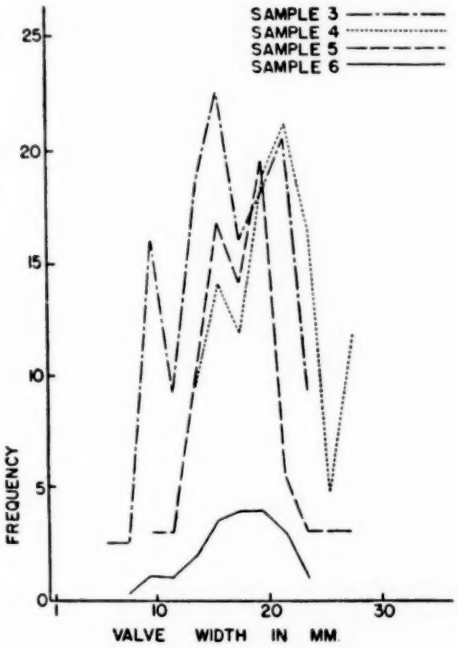

Fig. 8. Size-frequency distribution of *Antispirifer harroldi* (number of specimens per 10-cm cube).

found in them are apt to have belonged to the biocoenosis. These deposits will also be an excellent medium for the preservation of a thanatocoenosis composed of pelagic and planktonic organisms.

DEATH ASSEMBLAGES AND A LIFE ASSEMBLAGE FROM THE LOWER DEVONIAN OF NORTHERN MAINE

Certain of the variables considered above were used in the study of fossil assemblages from Lower Devonian strata in Somerset County, Maine. Most of the material studied came from six different beds in Parlin Gorge, about 2 miles northeast of the village of Lake Parlin, Parlin Pond Township, Somerset County, Maine.

About 95 per cent of the rocks exposed in Parlin Gorge

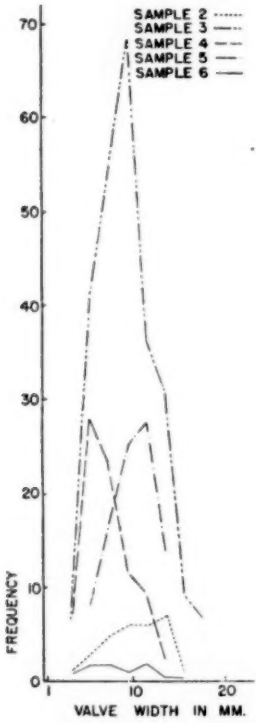

Fig. 9. Size-frequency distribution of *Leptocoelia flabellites* (number of specimens per 10-cm cube).

consist of medium-grained, dark blue-gray, argillaceous, brown-weathering, massive sandstone (subgraywacke). The remaining beds, scattered throughout the sequence, are slates. Fossils are ordinarily abundant in the slates, uncommon in the sandstone.

The ensuing paragraphs will attempt to analyze data from direct measurements of a number of specimens (table 1), taken from six samples of fossil-bearing slate and argillaceous sandstone, weighing 104 pounds. The fossiliferous samples were selected as being representative of the particular bed under investigation. All unfossiliferous rock adhering to the selected sample was removed in order to avoid a false picture of the population density, and the trimmed specimen was weighed. The weights of the samples ranged from 7¾ to 29 pounds. The weighed samples were then split parallel to the bedding, and records were kept of the number of valves belonging to each species and the maximum width of each valve.

TABLE 1

Number of Specimens Recovered from Six Fossil Beds in Parlin Gorge, Somerset County, Maine

Sample No.	1	2	3	4	5	6
Field No.	27	29	30	31	32	33
Antispirifer harroldi	...	...	59	50	24	92
"Spirifer" murchisoni	...	96	...	...	...	...
Leptocoelia flabellites	...	29	112	35	33	42
Mutationella sp.	...	176	64	1	2	2
Globithyris callida	231	...	...	...	...	...
f*	1.5	3.0	3.0	7.0	8.0	1.0
Sample weight in pounds	13.25	17.50	7.75	17.50	17.00	29.50

* "f" is the factor by which the observed values were multiplied to adjust for the estimated numbers of unobserved specimens in each sample.

It is not possible to count and measure some specimens in a given layer because they are packed so tightly that freeing one specimen breaks another, and some layers can be split only to a certain thinness. Therefore, an estimate "f" of the percentage of fossils not recorded was made.

Finally, these data were reduced to a common denominator, the number of specimens in a 10-cm cube, taking into account the estimated percentage of fossils not recorded and the varying weights of the initial samples. Specimens were measured to the nearest tenth of a millimeter, all measurements between each

odd millimeter and the succeeding even one being combined before reducing to a common denominator.

For five of the six beds studied there was a distinct development of a bell-shaped size-frequency distribution, whereas the size-frequency distribution in the sixth (sample 1) was strongly

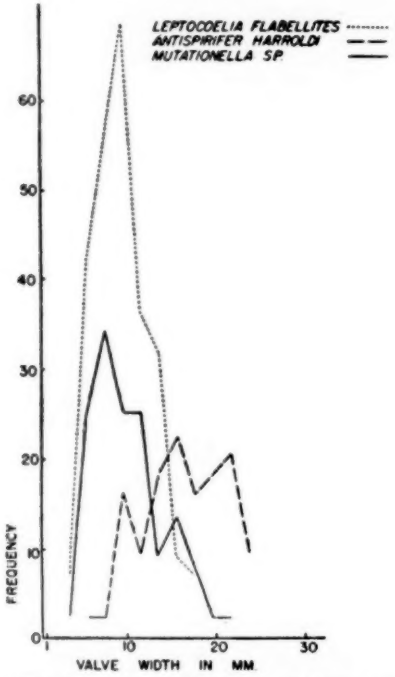

Fig. 10. Size-frequency distribution of fauna in sample 3 (number of specimens per 10-cm cube).

skewed to the right and concave upward (figs. 6-11). The curves also demonstrate that the size-frequency distribution of each species has its own mode in the same bed, presumably because each species has its own hydrodynamic properties and behaves in characteristic fashion when subjected to the same forces.

In curves for certain beds the modes are farther to the right than in others, but the bell-shaped outlines of the size-frequency distribution curves are about the same. Herein may lie a partial explanation for certain "dwarf" faunas, as previously suggested by Cloud (1948). The same mechanism also accounts for some

faunas composed mainly of large individuals, the Oriskany sandstone fauna of New York being an example.

All of the genera except *Leptocoelia* (fig. 6) are usually found as disarticulated shells. In none of the beds were many articulated shells found, with the exception of *Leptocoelia* and, in sample 1, of *Globithyris* (fig. 11). In other parts of Somerset County, *Globithyris* commonly occurs disarticulated. Sample 1 is composed of a rock rich in free carbon and iron sulfide.

The ratio of pedicle: brachial valves of the "*Spirifer*" (fig. 8) in sample 2 was approximately 3.5: 1.0.

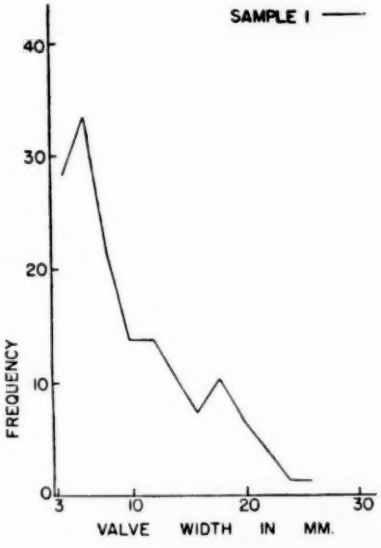

Fig. 11. Size-frequency distribution of *Globithyris callida* (number of specimens per 10-cm cube).

The population density of *Globithyris* in sample 1 for shells more than 10 mm wide is much less than that found in the probable death assemblages of this genus elsewhere in Somerset County, Maine. In general, the population density of sample 1 is lower than that of samples from the other beds in Parlin Gorge, indicating that these other beds may contain the remains of more sets than sample 1.

It is concluded that the fossils in samples 2 to 6 (figs. 6-10) probably represent death assemblages because of their size-

frequency distributions, the lack of articulated shells of those species which probably are easily disarticulated, and the ratio of opposite valves for "*Spirifer*" in sample 2, as well as the population densities. The beds represented by these samples are not very ferruginous or noticeably carbonaceous as compared with the bed that yielded sample 1. The fossils in sample 1 are thought to represent a probable life assemblage because there are numerous articulated shells of a species that is usually disarticulated with ease, the size-frequency distribution is strongly right-skewed, and the bed is rich in free carbon and iron sulfide.

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U. S. Geological Survey
Washington, D. C.

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THE PEGMATITIC DIFFERENTIATES OF BASIC SHEETS

FREDERICK WALKER

ABSTRACT. The coarse-grained veins, patches and schlieren found in so many dolerite[1] sills are considered in the light of recent work on crystallization differentiation. Chemical and petrographic data, some of them new, are given for examples taken from the Karroo, Palisadan and Whin Sill suites. It is shown that crystallization differentiation can account satisfactorily for most of the phenomena encountered but that there are two types of dolerite-pegmatite. The normal type is concordant and characterized by moderate iron enrichment. The other forms cross-cutting veins and shows extreme iron enrichment. The genesis of the two types is discussed.

INTRODUCTION

ALL geologists who have worked on dolerite sheets are familiar with the pegmatitic streaks, veins and patches enclosed by the normal rock. These have been given the name dolerite-pegmatite, since they possess the coarse grain and graphic intergrowth which are two main features of the pegmatites proper. Their bulk is always insignificant compared with that of the normal rock, but they are of importance to petrologists, for they provide clues concerning the crystallization and differentiation trends of igneous masses which are otherwise monotonously uniform. Furthermore, the sheets of dolerite containing them penetrate strata over vast areas of the earth's crust and are representative of the basaltic magma which dwarfs all others in volcanic associations.

The bearing of dolerite-pegmatites (and of the corresponding pegmatitic phases of basic lavas) on the differentiation problem has been emphasized by Shannon (1924), Lacroix (1928), Fenner (1929), and Kennedy (1933), but since the last of these publications much new information has accumulated on the problem of basaltic differentiation. In particular the magnificently exposed Skaergaard intrusion has been described in great detail by Wager and Deer (1939) who have thereby added much to our knowledge, not only of the crystallization process of basic magma, but also of the variation in

[1]The terms "dolerite" and "dolerite-pegmatite" used in this article, by courtesy of the editor, are the equivalents of the American terms "diabase" and "diabase-pegmatite."

the individual mineral series. In view of this, and of other recent contributions on similar subjects, the time seems ripe for a re-examination of the dolerite-pegmatites, especially from the genetic standpoint. The present article deals with examples from the states of Virginia, Pennsylvania, New Jersey and Connecticut in the Palisadan province; from the Whin Sill of northern England and other sills of the same intrusive episode in Central Scotland; and from the Karroo dolerite suite of the Union of South Africa.

These three provinces are representative of the tholeiite or saturated basalt magma-type. Olivine-dolerites do indeed have pegmatitic modifications, but sheets of this magma-type show a pronounced tendency to form composite intrusions and towards gravitational differentiation after emplacement. Thus the genetic problems are more complex in their case.

In order to avoid the danger of misinterpreting the work of others, all the examples described below have been studied by the author both in the field and in the laboratory. The American field work was carried out with the most helpful cooperation of Professor A. F. Buddington and other members of the Princeton geological faculty, of Professor A. Knopf of Yale University, and of Mr. F. Oldak of Philadelphia. For this valuable aid the author is sincerely grateful. His cordial thanks are also due to the Council of the University of Cape Town for a grant from the Staff Research Fund which defrayed the cost of the five new chemical analyses.

Finally, a tribute must be paid to C. N. Fenner's pioneer work on iron-enrichment which has had such a profound influence on all recent writings on basaltic crystallization (Fenner, 1929, 1931 and 1937).

FIELD OCCURRENCE OF DOLERITE-PEGMATITE

Dolerite-pegmatite is, in the main, a sill rock and the present article is confined to sills, since the existence of a definite roof and floor simplifies the differentiation problem. Similar pegmatitic modifications are found in lavas, but our knowledge of them is relatively scanty.

The three ways in which dolerite-pegmatite occurs in sills are as lens-shaped or tabular schlieren parallel to the contacts, as irregular nests or patches, and as cross-cutting veins often referred to as dikes.

Schlieren represent the most common and abundant mode of occurrence of dolerite-pegmatite. They are of stratiform or lenticular habit and have a maximum thickness of 2 m. (Whin Sill) in the examples considered. The average thickness is 15-60 cm. Exposures do not often permit an estimate of their maximum lateral extent, but a band was traced for 12 m. by Shannon (1924, p. 18) and one in the Whin Sill (quarry south of Middleton-in-Teesdale) was seen by the author to extend for at least 25 m. Individual schlieren may branch or give off ramifying veins (Tomkeieff, 1929, p. 103). Schlieren are usually confined to the upper half of sills though occasional exceptions have been noted, e.g., in the Hangnest sill (Karroo) table 1 (Walker and Poldervaart, 1941, p. 432).

Patches are often associated with schlieren in the upper portions of sills and differ from them in their irregular habit and relatively small lateral extent. They do not exceed 2 m. in length, the major diameter being roughly parallel to the contact.

Veins may occur at any horizon in sills and, though definitely cross-cutting, are rather irregular in their trend. In very few cases are they aligned parallel to any fracture system in the sill, and they generally follow a devious course in which the inclination is nearly always steep. The author has failed to trace any vein for more than 10 m. The veins are much thinner than the schlieren or patches, 20 cm. being the maximum width recorded and 1 cm. the minimum. In no case was a vein traced up to the contact.

General observations.—A feature of all three modes of occurrence is the great frequency of sharp, unchilled junctions with the host rock. Gradational contacts are rarer (Walker, 1940, p. 1063; Tomkeieff, 1929, p. 104) and chilled junctions absent. There is never any alteration of the adjacent dolerite by the dolerite-pegmatite.

The minerals of the dolerite-pegmatites, apart from graphic intergrowths, show either random orientation or have grown inwards from the contact which though sharp is often continuous. Some of these minerals, in particular bladed pyroxene, show marked curvature.

Other pegmatitic phases of dolerite sills.—All the masses of dolerite-pegmatite described are of small dimensions. At the most they form 5 per cent of the total thickness of the

sill in any cross section (5 per cent at Cauldron Snout and 3 per cent at Crook boring, both in Whin Sill) (Tomkeieff, 1929, pp. 103-104). The volume percentage of dolerite-pegmatite is very much less and probably never exceeds 1 per cent.

There are, however, other coarse-grained modifications in some of the thicker quartz-dolerite sills of the Midland Valley of Scotland which may form up to 25 per cent of certain cross-sections or bore-cores. They are of somewhat different habit and genesis from the dolerite-pegmatites mentioned above and occur as an upper coarse-grained phase of the normal dolerite into which they are gradational. They contain even more abundant micropegmatite than do normal dolerite-pegmatites but no graphic intergrowth between pyroxene and plagioclase. Both pyroxene and iron ore occur as peculiar branching forms.

Robertson (1937) has given a general account of the field characters of these sills and an example in Stirlingshire has recently been described by the author (Walker, 1952). Shannon (1924, pp. 22-23) has described a number of quartz-albite-diopside veins from Goose Creek quarry which appear to have been formed by the hydrothermal alteration of normal dolerite-pegmatite.

Dolerite-pegmatite presents a marked contrast to the light-colored aplitic or felsitic veins which often accompany them and present a problem akin to the aplite-pegmatite association of granite masses.

PETROGRAPHY OF DOLERITE-PEGMATITES

Dolerite-pegmatite contains practically the same range of minerals as normal dolerite with two exceptions: Orthopyroxene is completely absent and greenish-brown amphibole is generally a major constituent.

The chief differences between dolerite-pegmatite and the associated dolerite are in the relative proportions of the major constituents and in the fabric. The dolerite-pegmatites contain abundant micropegmatite but in contrast are also rich in iron ore (table 1). Both the plagioclase and pyroxene, which are the two main constituents, develop long, bladed forms often showing marked curvature. The relationship between these minerals is generally sub-ophitic except in varieties with an unusually large proportion of micropegmatite. In such cases both minerals are idiomorphic and there is a transition into

TABLE 1

Mode of Occurrence of Dolerite-pegmatites and Mineralogical Data

Locality and reference to chemical analyses	Thickness of sill (meters)	Position in sill	Habit	Percentage of pyroxene. β of augite.	Percentage of plagioclase. An content.	Percentage of iron ore	Percentage of quartz and micropegmatite	Percentage of all other constituents	Color index
Goose Creek, Va.	750?	Near	Schliere	32.0	44.0	5.0	13.0	6.0	43
Table 3, 9.		base		1.708	An_{55}				
Lambertville, N. J.	500	150 m	Vein	23.0	43.0	15.0	14.0	5.0	43
Table 3, 2.		above base		1.707	An_{45}				
		180 m	Schliere	32.3	48.2	6.9	6.0	6.6	44
		above base		1.705	An_{45}				
Palisades, N. J.	300	10 m	Vein	20.0	37.0	14.0	8.0	21.0	45
Table 3, 4 and 5.		above base		1.705	An_{48}				
West Rock, Conn.	100	5 m	Schliere	25.0	47.0	8.0	12.0	8.0	41
Table 3, 7.		below top		1.706	An_{52}				
Hangnest, Karroo.	150	30 m	Schliere	27.8	31.3	4.9	29.2	6.8	36
Table 3, 12.		above base		1.710	An_{30}				
			Lens	30.2	38.5	5.5	21.9	3.9	40
Alewyn's Gat, Karroo.	150	50 m		1.708	An_{50}				
Table 3, 13 and 14.		above base	Schliere	23.0	32.0	5.0	35.0	5.0	33
				1.707	An_{30}				
Libode (Transkei, S. Africa)	300	Middle	Schliere	26.0	45.0	10.0	16.0	3.0	39
Table 3, 16.				1.707	An_{50}				
Elandsberg (Natal, S. Africa)	90	4 m	Schliere	25.0	47.0	5.0	16.0	7.0	37
Table 3, 19.		below top		1.707	An_{55}				
High Force (Whin Sill)	50	Above	Schliere	23.0	37.0	13.0	18.0	9.0	46
(Tomkeieff, 1929)		middle		1.710	An_{48}				
Crook bore (Whin Sill)	55	Middle	Schliere	23.5	37.0	11.5	18.0	10.0	45
Table 3, 21.				1.710	An_{50}				

granophyre. Large skeletal crystals of iron ore are always conspicuous and biotite may become important locally.

Graphic intergrowths between quartz and sodic plagioclase, between quartz and orthoclase, and between calcic plagioclase and augite are exceedingly common and highly characteristic (plate 1).

Additional notes on individual mineral groups follow below:

Plagioclase in dolerite-pegmatites is nearly always slightly more sodic than that of the host dolerite. When zoning is present it is progressive but only conspicuous marginally. Portions showing graphic intergrowth with augite are separate from the main crystals, though often in optical continuity with them, and are of the same composition.

The sodic margins of crystals may extend outwards in optical continuity with quartz-orthoclase micropegmatite. Alteration of plagioclase to turbid or micaceous products is common and sometimes accompanied by albitization.

Pyroxene is chiefly purple or brown augite or ferroaugite with characteristic bladed habit, often curved. Some crystals show a conspicuous median line which forms the spine of typical herringbone twins. Colorless pigeonite, sharply defined but in optical continuity with the augite, forms the center of many columns but is so prone to alteration that it has frequently escaped detection.

A sheaf-like disposition of the augite blades is sometimes seen and the terminal portions of some may show an almost variolitic interleaving with plagioclase.

A good deal of the pyroxene has gone over to brownish-green amphibole, especially the marginal portions and the graphic inclusions.

Iron ore is a more abundant constituent of dolerite-pegmatite than of the associated dolerites and forms large, conspicuous crystals. It varies in form from platy to skeletal and has an extensive crystallization range. Polished sections viewed by reflected light show it to be an intergrowth of ilmenite with magnetite or hematite, with the addition of occasional scaly pyrite.

Micropegmatite is always abundant and either orthoclase or sodic plagioclase may act as the host mineral. The former host is generally turbid but the latter may be unaltered. The intergrowth varies from medium to coarse grain but may be

Table 2

New Chemical Analyses and Norms of Dolerite-pegmatites and Associated Dolerites

	Analyses						Norms				
	1	2	3	4	5		1	2	3	4	5
SiO_2	51.25	49.02	51.44	52.02	45.90	Q	1.5	7.3	3.7	8.6	3.6
TiO_2	0.96	3.94	1.32	1.52	5.98	or	3.3	6.1	3.3	3.3	5.6
Al_2O_3	16.69	13.13	16.26	15.66	9.36	ab	18.9	22.0	19.4	18.9	26.7
Fe_2O_3	0.61	5.66	1.74	3.85	8.02	an	33.9	21.1	32.3	31.0	8.5
FeO	9.84	13.57	8.58	8.38	10.39	di wo	7.3	5.5	7.8	6.5	13.7
MnO	0.22	0.20	0.22	0.21	0.28	di en	3.4	1.9	4.1	3.4	10.1
MgO	5.83	2.84	5.81	4.58	5.42	di fs	3.8	3.7	3.4	2.9	2.2
CaO	10.34	7.22	10.43	9.56	8.56	hy en	11.2	5.2	10.5	8.1	3.5
Na_2O	2.23	2.59	2.36	2.22	3.18	hy fs	12.5	10.4	8.8	7.1	0.8
K_2O	0.52	1.07	0.56	0.60	0.93	mt	0.9	8.1	2.6	5.6	11.6
H_2O+	0.96	0.45	1.09	1.02	1.12	il	1.8	7.5	2.6	2.9	11.4
H_2O-	0.34	0.24	0.28	0.43	0.57	ap	....	0.7	0.3	0.3	0.3
P_2O_5	0.05	0.26	0.08	0.15	0.15	H_2O	1.3	0.7	1.4	1.5	1.7
CO_2	nil	nil	nil	nil	nil						
Total	99.84	100.19	100.17	100.20	99.86		99.8	100.2	100.2	100.1	99.7

1. Dolerite from quarry W. of Lambertville, N. J., 150 m above base of sill. Analyst, W. H. Herdsman.
2. Dolerite-pegmatite vein cutting 1. Analyst, W. H. Herdsman.
3. Dolerite 5 m below upper contact sill, West Rock, Conn. Analyst, W. H. Herdsman.
4. Dolerite-pegmatite schlieren in 3. Analyst, W. H. Herdsman.
5. Dolerite-pegmatite vein cutting dolerite below olivine ledge of Palisade sill, George Washington Bridge, N. J. Analyst, W. H. Herdsman.

accompanied by a much finer and streaky form (e.g., at Lambertville, N. J.) which seems to replace calcic plagioclase and is surrounded by normal micropegmatite.

Amphibole is generally an important constituent and is never absent. It is pale brown or deep greenish-brown in color and occurs mainly as a parallel, marginal outgrowth from augite or as a replacement of graphic pyroxene in which case the optical continuity of the inclusions is maintained. A few scraps of dark-green or blue alkali-amphibole may occur associated with micropegmatite.

MICROMETRIC DATA

The micrometric data of table 1 bring out clearly the distinguishing characteristics of dolerite pegmatite. No figures are given for grain-size since the large dimensions of the constituent minerals render an accurate estimate impracticable. The micrometric percentages given are guides rather than reliable figures.

The minimum grain-size for dolerite-pegmatite is at least double that of the host dolerite, pyroxene attaining the largest dimensions. The author has often seen blades up to 8 cm. long and Shannon (1924, p. 15) has recorded a length of 20 cm. at Goose Creek.

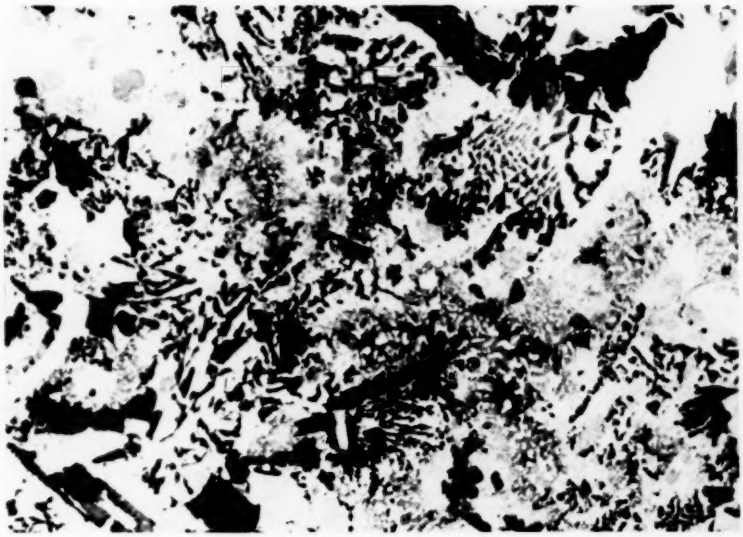
The mineral composition of dolerite-pegmatite is characterized by much higher proportions of micropegmatite and iron ore than in the normal dolerite. In the two cross-cutting veins in table 1 the proportion of iron ore is particularly great.

Although there is less pyroxene than in the host dolerite, the deficiency is largely counterbalanced by greater abundance of amphibole, so that the color indices of the two types are much the same. The plagioclase percentage is lower in dolerite-pegmatite and the average composition distinctly more sodic (An_{50} against $An_{60\text{-}65}$). Alkali feldspar is always more abundant in dolerite pegmatite than in the parent dolerite.

CHEMICAL DATA

The chemical analyses in tables 2 and 3 reflect, on the whole, the mineralogy of the dolerite-pegmatites. The silica percentage (46.8-58.6) shows a considerable overlap on either side of the

PLATE 1



Thin section of dolerite-pegmatite vein, Lambertville, N. J. × 2½. Ordinary light. Graphic intergrowths of alkali feldspar and quartz, and of pyroxene and plagioclase, are conspicuous throughout the field. A curved crystal of plagioclase is seen at right. Some of the pyroxene is replaced by dark amphibole.

values for dolerite (51.2-55.6) and the two lowest values are shown by the cross-cutting veins of the Palisade sill. Both these rocks carry normative and modal quartz due to the unsilicated condition of much of the iron oxide and titania.

Dolerite-pegmatite invariably shows lower lime and magnesia than the associated dolerite and this makes itself felt in the relatively iron-rich pyroxenes and more sodic plagioclase which characterize the coarser rock.

Ferric oxide, titania and potash are nearly always more abundant in dolerite-pegmatite than in the associated dolerite, this being reflected in the greater proportions of titaniferous iron ore and micropegmatite. The variation of the minor constituents is somewhat erratic.

The three important ratios $\frac{(FeO + Fe_2O_3) \times 100}{MgO + FeO + Fe_2O_3}$, $\frac{Fe_2O_3 \times 100}{FeO + Fe_2O_3}$, and $\frac{K_2O \times 100}{Na_2O + K_2O}$ are shown graphically in figure 1. The first ratio is the most reliable since the analytical errors in the estimation of MgO and total iron are relatively slight. It is also the most significant for it provides a measure of the differentiation of normal basaltic magma. It can be used accordingly as the abscissa of variation diagrams in which the oxides are taken as the ordinates. In this respect it is far superior to the silica percentage which tends to remain static in basaltic differentiation owing to a rough balance between the plagioclase and pyroxene series, the one showing increasing and the other decreasing silica with progressive differentiation.

The ratios $\frac{Fe_2O_3 \times 100}{FeO + Fe_2O_3}$ and $\frac{K_2O \times 100}{Na_2O + K_2O}$ are less reliable owing to greater possibility of analytical errors. Marked discrepancies between the analytical and petrographic data are noticed, for instance, in the analyses of 9 and 10, and 20 and 21 in table 3. In both the figures for Fe_2O_3 in the dolerite-pegmatite seem much too low. It is accordingly impossible to apply in such cases the important experimental results of G. C. Kennedy (1948) on the relationship between the volatile and ferric iron contents of igneous rocks.

TABLE 3

Analyses of Dolerite-pegmatites of Sheets and Associated Dolerites (in *italics*) Calculated 100% Anhydrous

	1	2	3	4	5	6	7	8	9	10
SiO_2	*52.0*	49.3	*52.3*	53.4	46.8	*51.5*	*52.0*	52.5	*51.5*	53.0
TiO_2	*1.0*	4.0	*1.4*	1.4	6.1	*1.4*	*1.3*	1.5	*1.5*	2.3
Al_2O_3	*16.9*	13.2	*15.0*	18.5	9.5	*14.1*	*16.4*	15.9	*13.8*	14.8
Fe_2O_3	*0.6*	5.7	*1.2*	2.0	8.2	*3.6*	*1.8*	3.9	*1.0*	0.2
FeO	*10.0*	13.6	*9.4*	7.7	10.5	*8.2*	*8.7*	8.5	*11.3*	12.0
MnO	*0.2*	0.2	*0.1*	0.1	0.3	*0.4*	*0.2*	0.2	*0.2*	0.2
MgO	*5.9*	2.9	*7.4*	2.6	5.5	*7.6*	*5.9*	4.7	*7.4*	5.4
CaO	*10.5*	7.3	*10.2*	9.4	8.7	*10.6*	*10.5*	9.7	*10.1*	8.3
Na_2O	*2.3*	2.6	*2.1*	3.8	3.2	*2.1*	*2.4*	2.3	*2.1*	2.0
K_2O	*0.5*	1.1	*0.8*	1.0	1.0	*0.4*	*0.6*	0.6	*1.0*	1.5
P_2O_5	*0.1*	0.1	*0.1*	0.1	0.2	*0.1*	*0.2*	0.2	*0.1*	0.3

TABLE 3 (*Cont.*)

	11	12	13	14	15	16	17	18	19	20	21
SiO_2 ..	*54.6*	58.6	*52.0*	53.2	56.9	*52.7*	53.6	*55.6*	56.0	*51.2*	50.7
TiO_2 ..	*1.2*	1.2	*0.9*	2.0	2.0	*0.8*	1.4	*1.3*	1.6	*2.4*	2.8
Al_2O_3 .	*16.6*	14.7	*15.4*	13.2	13.3	*12.2*	13.4	*15.9*	14.2	*14.8*	16.0
Fe_2O_3 .	*0.4*	2.6	*1.1*	2.2	3.6	*0.3*	2.0	*1.2*	1.8	*3.5*	2.5
FeO ...	*9.5*	8.4	*10.7*	10.7	10.4	*10.3*	11.3	*8.0*	8.5	*9.0*	11.6
MnO ..	*0.1*	0.1	*0.2*	0.2	0.2	*0.2*	0.4	*0.2*	0.2	*0.2*	0.1
MgO ..	*5.3*	3.2	*6.5*	5.2	2.3	*10.3*	5.8	*6.3*	5.9	*5.4*	3.1
CaO ...	*9.2*	7.3	*9.9*	9.5	6.7	*10.9*	8.6	*8.0*	8.4	*8.9*	8.2
Na_2O .	*2.1*	2.0	*2.5*	2.7	2.2	*1.7*	2.3	*2.9*	2.6	*2.6*	3.0
K_2O ..	*0.8*	1.6	*0.7*	0.9	2.0	*0.5*	0.9	*0.6*	0.8	*1.2*	1.2
P_2O_5 ..	*0.2*	0.3	*0.1*	0.2	0.4	*0.1*	0.3	*tr.*	tr.	*0.3*	0.6
etc. ...	...	...	...	...	...	...	...	...	...	*0.5*	0.2

1. Dolerite, 150 m above base of Lambertville sill, N. J. Analyst, W. H. Herdsman (new analysis).
2. Dolerite-pegmatite vein, adjacent to 1. Analyst, W. H. Herdsman (new analysis).
3. Basaltic dolerite, 25 m above base of Palisade sill, George Washington Bridge, N. J. Analyst, F. A. Gonyer, quoted from Walker, 1940, table 3.
4. Dolerite-pegmatite schliere, 180 m above base of Palisade sill. Analyst, W. H. Herdsman, quoted from Walker, 1940, table 3.
5. Dolerite-pegmatite vein, Palisade sill, 10 m above 3. Analyst, W. H. Herdsman (new analysis).
6. Dolerite, West Rock sill, Conn. Analyst, G. W. Hawes, quoted from Shannon, 1924.
7. Dolerite, 5 m below upper contact, West Rock sill, Conn. Analyst, W. H. Herdsman (new analysis).
8. Dolerite-pegmatite schliere adjacent to 7. Analyst, W. H. Herdsman (new analysis).

9. Dolerite near base of Goose Creek sill, Va. Analyst, E. V. Shannon, quoted from Shannon, 1924, p. 13.
10. Dolerite-pegmatite vein near 9. Analyst, E. V. Shannon, quoted from Shannon, 1924, p. 19.
11. Basalt, lower contact of Hangnest sill, Karroo. Analyst, W. H. Herdsman, quoted from Walker and Poldervaart, 1949, table 16.
12. Dolerite-pegmatite schliere, 30 m above base of Hangnest sill. Analyst, W. H. Herdsman, Walker and Poldervaart, 1949, table 16.
13. Dolerite, 30 m below upper contact of sill, Alewyn's Gat, Karroo. Analyst, W. H. Herdsman, Walker and Poldervaart, 1949, table 16.
14. Dolerite-pegmatite, lens, near 13. Analyst, W. H. Herdsman, Walker and Poldervaart, 1949, table 16.
15. Dolerite-pegmatite schliere below xenolith near 13. Analyst, W. H. Herdsman, Walker and Poldervaart, 1949, table 16.
16. Dolerite, middle of sill near Libode, Karroo. Analyst, W. H. Herdsman, Walker and Poldervaart, 1949, table 16.
17. Dolerite-pegmatite schliere, adjacent to 16. Analyst, W. H. Herdsman, Walker and Poldervaart, 1949, table 16.
18. Dolerite, 30 m above lower contact of 100 m sill, Elandsberg, Natal. Analyst, W. H. Herdsman, Walker and Poldervaart, 1949, table 16.
19. Dolerite-pegmatite schliere, 26 m above 18. Analyst, W. H. Herdsman, Walker and Poldervaart, 1949, table 16.
20. Average dolerite of Whin Sill. Quoted from Tomkeieff, 1929, table 3.
21. Dolerite-pegmatite schliere, Crook bore, Whin Sill. Analyst, W. H. Herdsman, quoted from Tomkeieff, 1929, table 3.

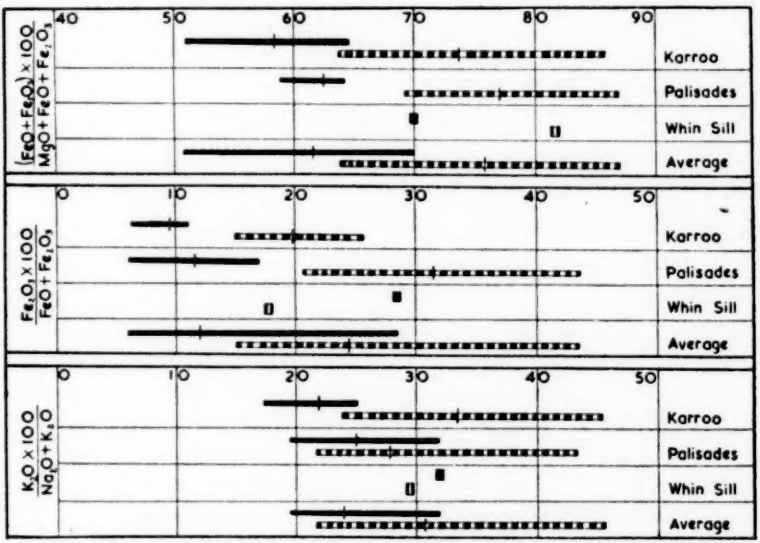


Fig. 1. Ratios of total iron to magnesia; ferric iron to total iron and potash to total alkalies. Pecked line indicates dolerite-pegmatite and unbroken line dolerite. Short vertical line shows average for province.

The results, plotted on figure 1, show an increase from dolerite to dolerite-pegmatite for all three ratios. In the few exceptions noted the analytical results are definitely suspect, and the general increase is in keeping with the differentiation trend of all normal basaltic provinces which have been studied in detail.

GENESIS OF DOLERITE-PEGMATITE

Previous work.—Both Shannon (1924) and Tomkeieff (1929) have considered the genesis of dolerite-pegmatite and recognized that this coarse-grained rock-type, with its hydatogenic mineralogy, crystallized from volatile-rich fractions of normal dolerite magma. But, whereas Shannon (1924, p. 39) remained agnostic as to the mechanism whereby such volatile-

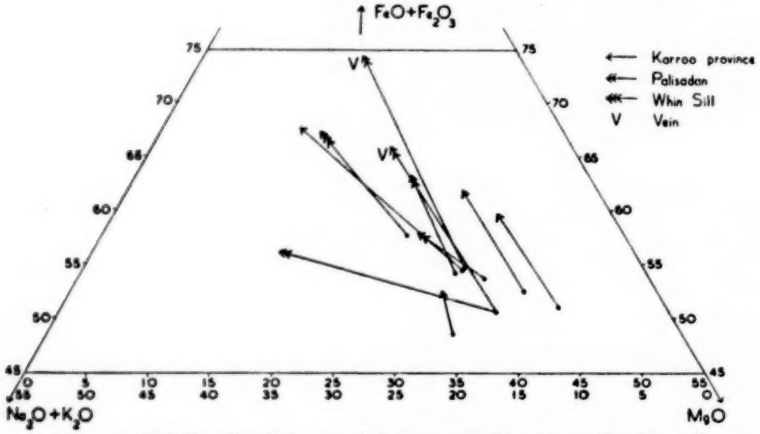


Fig. 2. Relationship between dolerite-pegmatite (spear-head) and parent dolerite (spear-butt).

rich fluids were concentrated, Tomkeieff (1929, p. 114) considered that segregation took place prior to emplacement and represented a kind of pre-injection liquation differentiation.

Position of dolerite-pegmatite on differentiation-trend.—Recent work has tended to support the views of Shannon (Edwards, 1942, pp. 477-480; Walker and Poldervaart, 1949, pp. 662-663). In every case for which data are available dolerite-pegmatite lies further along the differentiation path than the parent rock. This is fully confirmed by the present investigation and in figure 2, which is a modification of Walker

and Poldervaart's diagram (1949, fig. 36) to include Fe_2O_3, a uniform variation trend towards iron enrichment is conspicuous. The iron-enrichment of the two cross-cutting veins (V) of dolerite-pegmatite is especially pronounced and the process, if continued under suitable physical conditions, might lead to the dense ore-rich liquids postulated by Bateman (1951).

In figure 3 the dolerite-pegmatites lie on the continuation towards the iron co-ordinate of the differentiation trend of

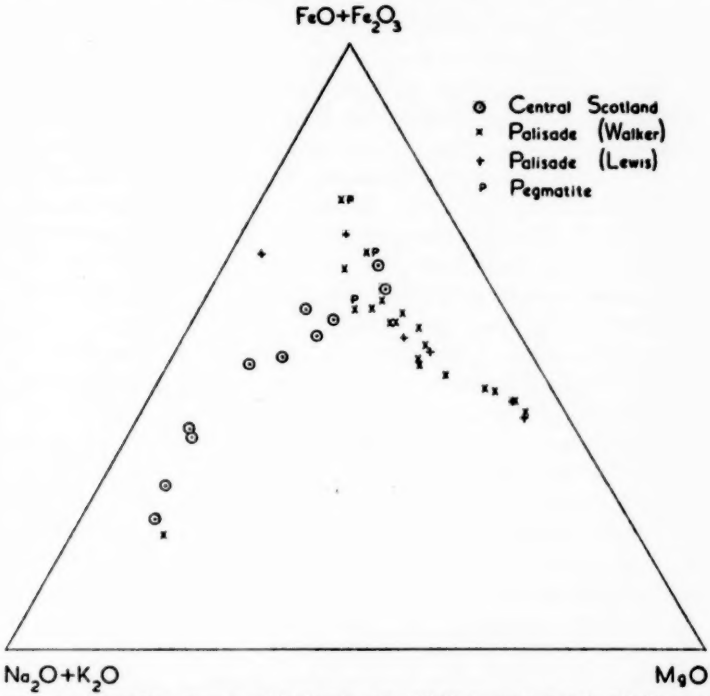


Fig. 3. Variation trends of Palisade sill (Walker, 1940, Lewis, 1937) and of quartz-dolerite sills of Central Scotland (Walker, 1952).

the Palisade sill, showing clearly that the pegmatitic types are linked genetically to those of normal crystallization. This matter will be discussed further at a later stage.

Mechanism of formation.—The question of the mechanism of segregation and injection of dolerite-pegmatite remains open, though the evidence is against the pre-injection liquation

differentiation theory suggested by Tomkeieff. Clues are afforded, however, by the habit and distribution of dolerite-pegmatites in sills and by their variation in chemical composition. Before these clues are followed it is necessary to consider the possibility of some dolerite-pegmatites arising through syntexis with silicic sediments. Only two analyses (nos. 12 and 15 of table 3) are suspect. Both are from the Karroo province where dolerite intrusions react readily with the associated sediments. The question of sedimentary syntexis in these two cases has been discussed at length elsewhere (Walker and Poldervaart, 1942, pp. 301-304) and the conclusion reached that it was unlikely.

The distribution of dolerite-pegmatites in sills seems to bear a certain relationship to the grain-size of the parent dolerite. Shannon (1924, p. 10) recorded at Goose Creek that dolerite-pegmatite is absent where the grain-size of the normal dolerite is large. In the Palisade sill, however, numerous schlieren of dolerite-pegmatite are found in the coarse-grained upper portion. It is perhaps noteworthy that in the relatively thin and fine-grained Whin Sill the proportion of true dolerite-pegmatite is greater than in any other example listed above. It seems possible that failure to segregate the volatile constituents and late minerals may cause a general instead of a local increase in grain-size.

Poldervaart (1944, p. 106) has suggested that dolerite-pegmatite is scarce or absent in basic intrusions showing a marked degree of iron-enrichment, e.g., those of the Skaergaard and New Amalfi. But these intrusions are cross-cutting and large in comparison with the sills now under consideration so that conditions were more favorable for convectional circulation of the magma and correspondingly unfavorable for segregation.

Sills in which dolerite-pegmatite is found do not seem to be over 500 m. thick and the depth of cover was probably of the order of 1 km. to 5 km. Convectional circulation is unlikely to be established in such circumstances, for not only would cooling be relatively fast, but the increase in melting point of the constituent minerals from roof to floor would be negligible. Furthermore the magmas under consideration are four to six times richer in potash, and therefore probably more viscous, than that of the Skaergaard intrusion, which provides the

only case with clearly demonstrable convection circulation (see discussion in Wager and Deer, 1939, pp. 266-270).

We are dealing, then, with a series of stratiform sheets, of great lateral extent but limited thickness, which cooled from both roof and floor without appreciable convectional circulation. The loss of heat from roof and floor was probably roughly equal for the upper and lower chilled phases of such sills are of similar dimensions when they are measurable. Formation of crystals continued inwards from the welded contacts though we do not know the details of the process. It is suggested that some crystals grew attached to the existing framework while others remained discrete to form a cloud, or even a mush, adjacent to it. The crystal framework was probably spongy at first with interstitial liquid containing potential late minerals.

The early discrete crystals of pyroxene (and sometimes iron ore and olivine as well) tended to sink in the magma and in thick sills of relatively low viscosity (e.g., the Palisade sill) the settling process became important (Walker, 1940, pp. 1083-1089). In the much thinner Whin Sill crystal settling was negligible. It is also possible that portions of the upper crystal framework became stoped off and sank in the magma, their large size compensating for a comparatively low specific gravity intermediate between that of calcic plagioclase and pyroxene.

The result of continued undisturbed crystallization was the formation of a layer of residual liquid constantly diminishing in thickness but increasing in volatile content until solidification was complete. Without appreciable crystal settling the layer of residual liquid coincided approximately with the median phase of the sill, but if gravitational differentiation was significant, the layer became situated above the median plane, its final position being determined by the amount of crystal accumulation. Such conditions, with only slight crystal settling, seem to have obtained in many of the quartz-dolerite sills of Central Scotland which have a thickness below 100 m. (Robertson, 1937; Walker, 1952). They contain a coarse-grained phase rich in micropegmatite which is transitional above and below into normal dolerite and is situated slightly above the median plane of the sill.

The development of dolerite-pegmatite demanded somewhat

different conditions for, whether concordant or cross-cutting, the type is sharply defined and does not often show gradational contacts. It represents variation along the liquid line of descent towards iron enrichment and forms the penultimate rest-magma, rich in iron and volatiles, of sheets in which crystallization is far advanced. This liquid, which was certainly of small amount and probably of low viscosity, must have been squeezed into rifts in the crystal mesh or entrapped as pools by the local union of unevenly growing upper and lower frameworks. Tyrrell (1928, pp. 565-566), when dealing with residual liquids of alkaline olivine-dolerite sills in Ayrshire, suggested a process of lateral squeezing of the rest-magma into rifts in the crystal mesh accompanied by a breaking down of the framework to yield trachytoid textures. He considered that the downward pressure exerted by the superincumbent rock column was an adequate motive power to produce these results. But Holland (1897, pp. 410-411), Holmes (1936, p. 231) and others have emphasized the strength of an augite-plagioclase framework and have pointed out that it would only collapse under extreme pressure. It would seem accordingly that filter-press action is not a factor in the formation of dolerite-pegmatite.

The occurrence of planes of weakness in a spongy crystal structure with residual liquid, or even in solid dolerite, is another matter and appears to be the only assumption capable of explaining the observed facts.

Horizontal rifts in a crystal framework, continuous locally from top to bottom of the sill, may have arisen by contraction in the manner suggested by Tyrrell (1928, pp. 565-566). If the residual liquid was forced laterally into these by continued crystallization, or was entrapped in pools as postulated above, an explanation is provided for the habit and distribution of patches and schlieren of dolerite-pegmatite. Such represent a liquid which has suffered only moderate differentiation. The cross-cutting veins belong to a later stage when there was less and more highly differentiated residual liquid. They represent indeed the extreme in iron enrichment of normal dolerite magma and were expressed as thin cross-cutting veins which made their way outwards through the almost completely solid sill. Edwards (1942, p. 480) suggests that they follow joint

planes, but the distribution noted by the author is not allied to any regular fracture system.

As the last stage of consolidation, the residual liquids, now of minute amount, became granophyric in composition through crystallization of abundant iron ore, and so enriched in water that hydrothermal reactions came into play. This led to the development of quartz-albite-diopside veins described by Shannon (1924, pp. 22-28) and Walker (1940, pp. 1093-1096). Shannon has described in detail varieties from dolerite-pegmatite to the true hydrothermal types and transitions from dolerite to granophyre have been recorded from the Karroo province by Walker and Poldervaart (1940, pp. 165-166).

In some cases extreme differentiation seems to have developed concordant bands of granophyric material within the sill as for instance in the very thick sill of Dillsburg, Pa. (Hotz, 1950). In such cases hydrothermal reaction may be slight so that the differentiation trend is carried on from the peak of iron enrichment towards the alkali corner of the diagram figure 3. In the Palisade sill itself the regularity of the variation is upset beyond the peak by hydrothermal reactions which cause scattering of the points (Walker and Poldervaart, 1949, fig. 30).

Relationship of chemistry to mineralogy.—The chemical analyses of dolerite-pegmatite show the result of the piling up of TiO_2 and iron oxide (in particular Fe_2O_3) in the magma by the early crystallization of minerals poor in or devoid of these constituents. The percentages given by Wager and Deer (1939, p. 311) as the minima for the primary precipitation of magnetite and ilmenite from basalt magma are 3.5 Fe_2O_3 and 2.5 TiO_2, both of which are exceeded in most analyses of dolerite-pegmatite. It is not therefore surprising that both these minerals should be conspicuous in the examples under investigation. The percentage of total iron plus TiO_2 in the two cross-cutting veins (nos. 2, 5, table 3) is no less than 25. It is only the unsilicated condition of these constituents in the form of iron ore that allows the crystallization of abundant quartz in rocks with silica percentages so low as 49 and 47. The building up of K_2O in dolerite-pegmatites is due principally to the limited entry of K^+ into the plagioclase structure.

The scattering of the points after the peak of iron enrichment has been passed, has been already noted in the case of the Palisade sill and similar features and scattering take place

in the points representing the Karroo analyses in the diagram of Walker and Poldervaart (1949, fig. 27), though here the problem is further complicated by the possibility of syntexis with sediments.

Quartz-dolerite sills of Central Scotland.—In marked contrast the differentiated quartz-dolerite sills of Central Scotland show in figure 3 a continuous and clear-cut trend from the initial magma, which lies close to the peak of iron enrichment, towards the alkali corner. Hydrothermal reactions certainly took place in these sills but they appear to have caused little alteration in the bulk chemical composition of the differentiates (Walker, 1952, p. 403).

In these sills the coarse phase above the median plane probably corresponds to the dolerite-pegmatite of other intrusions from which it is distinguished by much greater relative abundance and by differences of mineralogy and chemical composition originating in the initial magma. The average quartz-dolerite of Central Scotland contains (anhydrous) 2.6 per cent TiO_2 and 5.5 per cent Fe_2O_3 (Walker, 1935, table 4) so that primary precipitation of magnetite and ilmenite was not only possible but is strongly indicated by the habit of the iron ore in thin section. The early elimination of unsilicated iron oxides and titania built up SiO_2 in the rest-magma to a much greater extent than in normal dolerite-pegmatite and also lowered the content of ferromagnesian silicates so that the resultant coarse phase was distinctly granophyric in character. This phase displayed in addition a mobility and tendency to react with the earlier solid phases which calls to mind Fenner's (1938) account of basalt-rhyolite exposures of Gardiner River.

The high proportion of this coarse granophyric phase seems due not only to silication of the rest-magma by liberation of early iron ore but also to the relatively high alkali and low lime and magnesia of the original magma. The initial magma of the Whin Sill is close in chemical composition to that of the co-magmatic Scottish sills but contains considerably less Fe_2O_3 (Walker, 1935, table 4). It is not surprising therefore that the mode of differentiation of this intrusion is intermediate in many ways between that of the Scottish sills and of the Palisadan and Karroo examples described above.

Plagioclase-pyroxene intergrowths.—Since the composition of the late rest-magma is dependent on the relative rates of crys-

tallization of the earlier mineral series, the graphic intergrowths of plagioclase and pyroxene which are a feature of the dolerite-pegmatites are important in providing definite information under this head. In the Lambertville vein, which contains particularly fresh and abundant intergrowths, careful measurements of the ratio of pyroxene to plagioclase gave consistent results around 35:65 by volume, 38:62 by weight. The graphic stage of crystallization is fairly late in the range and it is interesting to find that the relative proportion of the two mineral series shows little change from that found in the earlier crystallates.

GENERAL CONCLUSIONS

Although the author's observations are limited in scope, they do not seem to be contradicted by published work dealing with the dolerite-pegmatites of other provinces. The following general conclusions have been reached:

(1) Dolerite-pegmatite in basic sheets represents a post-injection differentiate of the normal rock forced into rifts in the crystal mesh when crystallization was far advanced.

(2) Dolerite-pegmatite lies on the normal differentiation trend of dolerite magma on the liquid line of descent and is close to the stage of maximum iron enrichment.

(3) Veins of dolerite-pegmatite belong to a later stage of differentiation than patches or schlieren and may show extreme enrichment in iron and titania.

(4) The nature of the late differentiates of dolerite sills is largely controlled by differences of composition in the initial magma.

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CLAY MINERALS IN SEDIMENTARY ROCKS AND DERIVED SOILS

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ABSTRACT. Analyses of 43 clay fraction samples from sedimentary rocks and their derived soils, together with a review of published studies of soils developed on sedimentary material, reveal: (1) Clay minerals in soils formed on sedimentary rocks commonly are inherited from parent material. Some of the kaolinite-rich soils of southeastern United States have inherited the clay mineral from Coastal Plain deposits; at least some of the montmorillonite in soils of the western Great Plains has been inherited from montmorillonite-rich parent rock; illite is a common mineral in soils developed on illite-rich Pleistocene glacial deposits and Paleozoic marine shale. (2) Because illite is abundant in many kinds of sedimentary materials, this clay mineral group is commonly the predominant type in soils formed on sedimentary bedrock. (3) Alluvium carried to depositional basins contains considerable illite eroded from widespread, illite-bearing soils. (4) The composition of modern illite-rich alluvium deposited in the sea has not been significantly altered by the marine environment. (5) The kind of illite common in soils and alluvium generally has a lower K_2O content and SiO_2/Al_2O_3 ratio than the illite in many sedimentary rocks, indicating that inherited illite generally has been altered somewhat by soil-forming processes.

INTRODUCTION

BECAUSE soils were the main source of the clay fraction of modern alluvium and marine deposits and of sedimentary rocks, the clay mineral composition of soils and the origin of these minerals are an important aspect of the study of sediments.

In most of the general discussions of soil composition the abundant clay minerals present are considered to be primary products of weathering, the kind produced being controlled mainly by environmental conditions. Minerals of the kaolinite group predominate in soils of warm, humid regions; minerals of the montmorillonite and illite[1] groups are prevalent in the soils of cooler or drier regions. Inasmuch as kaolinite and montmorillonite apparently are the more common authigenic clay mineral groups, it has generally been assumed that illite is rare in soils and that the alluvium carried to depositional basins contains essential proportions of kaolinite and montmorillonite (Ross, 1945, pp. 182-183).

[1] In older publications kaolinite and montmorillonite were considered to be the only clay mineral types in soils.

The belief that kaolinite and montmorillonite are the most abundant and widespread clay minerals in soils is no longer tenable. Since illite was first reported as the principal mineral in some soil colloids by Hendricks and Fry (1930) numerous examples of illite-rich soils have been described. In fact, present evidence indicates that, although long and intense weathering may alter the composition of illite, an illite-type mineral nevertheless occurs abundantly in the regolith. Recently accumulating evidence also invalidates the assumption that the clay in soils developed on sedimentary rock is predominantly of primary origin. It should be clear, however, that the clay in soils developed on igneous and metamorphic rocks must be primary.

In a few of the many detailed studies of soil composition the clay minerals have been interpreted as inherited from parent sedimentary rocks. The clay in the montmorillonite-rich White Store soil (41)[2] of North Carolina, for example, is believed to have been derived from the montmorillonite-rich Triassic shale on which the soil has formed. The Norfolk soils (27) of North Carolina are developed on Cenozoic sediments of the Atlantic Coastal Plain, and in the colloidal fraction of both the soil and parent material there is a molecular SiO_2/Al_2O_3 ratio of kaolinite. "It is perhaps reasonable to assume that the greater part of the clay content now present in these soils existed in the soil parent material at the time it was laid down" (Holmes, Hearn, and Byers, 1938, p. 27). Most of the clay in a number of illite-rich Solonetz and Black Alkali soils probably was inherited from parent alluvial deposits (Kelley, Dore, and Page, 1941, pp. 121, 123). Similarly, the abundance of kaolinite in the Gosport soil (13) of Iowa has been attributed to its kaolinite-rich parent Pennsylvanian shale, and a study of illite-bearing Miami and related soils (25) in Ohio led to the conclusion that these soils may have inherited much of their clay from their parent material, illite-bearing Late Wisconsin glacial drift. Numerous other recorded analyses, such as those of Iowa soils by Peterson (1946) and by Russell and Haddock (1941) and those of Minnesota soils by Caldwell and Rost (1942), reveal no appreciable change in colloidal composition or clay mineral content with depth in the profile of soils developed on sedimentary rocks.

[2] Soils with references listed by number in table 1.

Despite evidence of this sort, inherited clay minerals in soil have not been extensively investigated, nor has this mode of origin been recognized as an important factor in most discussions of soil composition. The possibility of the derivation of soil clay from clay-bearing sedimentary material is of particular significance when considered in light of the fact that about 75 per cent of the soil of the United States has developed on sedimentary deposits and that illite minerals are known to be abundant in most shales, marine sedimentary rocks and modern marine sediments, as well as in loess and glacial drift, and in some underclay and gumbotil (see Grim, Lamar, and Bradley, 1937; Grim, Bray, and Bradley, 1937; Hursh, Lamar, and Grim, 1944; Grim, Dietz, and Bradley, 1949; Grim, 1951; and Weaver and Bates, 1950).

If soils formed on the widespread mantle of sedimentary material can inherit their clay minerals and if these minerals are relatively stable, then illite should occur more commonly in soils than is generally suggested in the literature, and it should be among the more common clay minerals carried by streams to depositional basins. The present study is an investigation of these possibilities by (A) assembling and discussing some of the many recorded analyses of soils and parent sedimentary material that bear on this problem, and by (B) determining the predominant clay minerals present in a varied collection of soils and their parent sedimentary rocks. This project has been supported by a research contract between the Office of Naval Research and Princeton University (contract N8onr-73800). The Department of Geology, Princeton University, generously provided transportation and laboratory facilities. The help of M. Lougheed, Wm. Poole, A. Sutherland-Brown, and Wm. Yahn as research assistants is gratefully acknowledged. It is a pleasure to record my indebtedness to Professor A. F. Buddington for his continued interest and helpful advice throughout the course of this investigation.

A

In the following discussion of clay minerals in the colloidal fraction of soil profiles the C horizon is considered to be parent material not yet affected by soil-forming processes. Similarity of composition of A and C horizons is emphasized here, but this is not to deny that there is significant movement or frac-

TABLE 1

Chemical Data and Clay Mineral Content of Colloidal Fraction of Soils and Sediments Discussed in Text

A. A horizon of soil
B. B horizon of soil
C. Parent material of C horizon

I — Illite group
M — Montmorillonite group
K — Kaolinite group

Soil		%K_2O	molecular SiO_2/Al_2O_3	Predominant clay mineral	Reference (Tech. Bulls. cited are of U. S. Dept. Agr.)
I Zonal and Intrazonal soils, Lithosols, and Palcosols					
1.	Badland silt loam, Lithosol	A. 2.70	A. 5.53		Tech. Bull. 502
	Interior, Jackson Co., South Dakota	B. 2.72	B. 5.37		This paper, table 2;5
	Oligocene White River formation (non-marine)	C. 2.61	C. 5.76	C. Illite and Montmorillonite	
2.	Barnes loam, Chernozem soil	A. 1.70	A. 4.07		Tech. Bull. 484
	Labolt, Grant Co., South Dakota	B. 1.44	B. 4.29	B. 40%I, 30%K, 20%M	Alexander, *et al.*, 1939
	Wisconsin glacial drift (calcareous)	C. 1.40	C. 4.41		
3.	Bethel silt loam, Planosol	A. 1.31	A. 3.63		Tech. Bull. 834
	2.5 miles south of Green Fork, Wayne Co., Ind.	B. 1.10	B. 3.40	B. 65%I, 25%M, 10%K	
	Wisconsin glacial drift (calcareous)	C. 3.30	C. 3.40		
4.	Brookston silty clay loam, Half-bog soil	A. 2.62	A. 3.41		Tech. Bull. 834
	Fairmont, Grant Co., Ind.	B. 2.49	B. 3.91	B. 65%I, 25%M, 10%K	Nelson and Hendricks, 1943
	Late Wisconsin glacial drift	C. 4.86	C. 3.27		
5.	Carrington loam, Prairie Soil	A. 1.44	A. 3.33		Tech. Bull. 484
	Winthrop, Buchanan Co., Iowa	B. 1.85	B. 3.17	B. 40%I, 40%K, 10%M	Alexander, *et al.*, 1939
	Iowan glacial drift (calcareous)	C. 1.89	C. 3.23		
6.	Cayucos sandy clay loam, Prairie soil	A. 0.66	A. 4.93	A. Illite	Barshad, 1946
	Rodeo, Contra Costa Co., Calif.	.....	.....		
	Unconsolidated non-calcareous sandstone	C. 0.94	C. 5.25	C. Illite	
7.	Decatur clay loam, Red and Yellow Podzolic soil	A. 1.24	A. 2.27		Alexander, *et al.*, 1939
	Russelville, Franklin Co., Ala.	B. .88	B. 2.12	B. 70%K, 10%I	Pearson and Ensminger, 1949
	Paleozoic massive limestone (marine)	C. .68	C. 2.02	C. Kaolinite	
8.	Dodgeville silt loam, Gray-Brown Podzolic soil	.....	.....	A. K-bentonite type, some Illite	Jackson, *et al.*, 1948
	Southern Wisconsin				
	Wisconsin glacial drift	.....	.....	C. Illite	
9.	Eutaw clay, Rendzina	.....	.....	A. M and K	Pearson and Ensminger, 1949
	Black belt, central Ala.	.....	.....		
	Upper Cretaceous marine deposits	.....	.....	C. M and K	
10.	Frederick silt loam, Gray-Brown Podzolic soil	A. 1.00	A. 2.24		Tech. Bull. 678
	Rockbridge Co., Va.	B. 1.22	B. 2.10	B. 70%K, 10%I	Alexander, *et al.*, 1939
	Paleozoic limestone and shale (marine)	C. 1.24	C. 2.10		
11.	Fullerton gravelly loam, Red and Yellow Podzolic soil	A. 0.86	A. 3.07		Tech Bull. 678
	Lawrence, Cherokee Co., Ala.	B. 1.28	B. 3.11	B. 60%K, 20%I	Alexander, *et al.*, 1939
	Paleozoic cherty limestone (marine)	C. 1.48	C. 2.61		
12.	Gleason clay loam, Prairie soil	.....	A. 3.77	A. Illite	Barshad, 1946
	12 miles northeast of Alturas, Modoc Co., Calif.	.....	.....		
	Cenozoic acid andesite tuff (non-marine)	.....	C. 3.59	C. Illite	
13.	Gosport silty clay loam, Gray-Brown Podzolic soil	.....	.....	A. Kaolinite	Peterson, 1946
	Marion Co., Iowa	.....	.....		
	Pennsylvanian Des Moines shale	.....	.....	C. Kaolinite	
14.	Hagerstown silty clay loam, Gray-Brown Podzolic soil	A. 1.00	A. 2.19		Tech. Bull. 678
	Center Co., Penn.	B. 1.45	B. 2.16	B. 70%K, 20%I	Alexander, *et al.*, 1939
	Paleozoic massive limestone (marine)	C. 1.24	C. 2.17		
15.	Hagerstown silt loam, Gray-Brown Podzolic soil	A. 1.51	A. 2.33		Tech. Bull. 678
	Hagerstown, Washington Co., Md.	B. 2.30	B. 2.44	B. 50%K, 40%I	Alexander, *et al.*, 1939
	Paleozoic massive limestone (marine)	C. 2.30	C. 2.45		
16.	Hays silty clay loam, Chernozem soil	A. 2.30	A. 4.25		Tech. Bull. 502
	1 mile south of Hays, Ellis Co., Kan.	B. 2.11	B. 4.29	B. K-bentonite type	Tech. Bull. 316
	Upper Cretaceous Carlisle shale (marine)	C. 2.11	C. 4.37		
17.	Hillsdale fine sandy loam, Gray-Brown Podzolic soil	A. 1.95	A. 3.02		Tech. Bull. 834
	Ingham Co., Michigan	B. 2.58	B. 2.94	B. 75%I, 10%K, 10%M	Nelson and Hendricks, 1943
	Late Wisconsin glacial drift	C. 3.41	C. 3.07		
18.	Houston black clay, Rendzina	A. 1.13	A. 3.90	A. Montmorillonite	Hendricks and Fry, 1930
	2 miles south of Temple, Bell Co., Texas	B. .99	B. 3.91		Tech. Bull. 316
	Upper Cretaceous Taylor marl (marine)	C. 1.07	C. 3.97		
19.	Houston clay, Rendzina	.....	.....	A. M and K	Pearson and Ensminger, 1949
	Black belt, central Ala.	.....	.....		
	Upper Cretaceous Selma chalk (marine)	.....	.....	C. M and K	
20.	Keith silt loam, Chestnut soil	A. 2.42	A. 4.61		Tech. Bull. 502
	Dundy Co., Nebraska	B. 2.16	B. 4.53		Swineford and Frye, 1951
	Pleistocene Peorian loess (non-marine)	C. 2.21	C. 4.52	C. M and I?	
21.	Lebanon silt loam, Red and Yellow Podzolic soil	A. 1.48	A. 3.38		Tech. Bull. 678
	St. James, Phelp Co., Mo.	B. 1.29	B. 3.29		
	Paleozoic cherty limestone (marine)	C. 0.79	C. 2.70		
22.	Marshall silt loam, Prairie soil	A. 2.14	A. 3.69		Tech. Bull. 316
	9 miles west of Clarinda, Page Co., Iowa	B. 1.58	B. 3.72	B. K-bentonite type	
	Pleistocene loess (non-marine)	C. 1.69	C. 3.57		
23.	Maury silt loam, Red and Yellow Podzolic soil	A. 1.90	A. 2.58		Tech. Bull. 678
	Ashwood, Maury Co., Tenn.	B. 2.42	B. 2.11		This paper, table 2; 30 and 31
	Paleozoic phosphatic limestone (marine)	C. 3.93	C. 3.24		
24.	Miami soil, Gray-Brown Podzolic soil	A. 2.42	A. 3.23		Tech. Bull. 834
	Williamsburg, Wayne Co., Ind.	B. 2.44	B. 3.17	B. 65%I, 25%M, 10%K	
	Late Wisconsin glacial drift (calcareous)	C. 3.56	C. 3.21		
25.	Miami soil, Gray-Brown Podzolic soil	.....	.....	A. Illite	Bidwell and Page, 1951
	Clark Co., Ohio	.....	.....		
	Late Wisconsin glacial drift	.....	.....	C. Illite	
26.	Mifflinburg soil, Gray-Brown Podzolic soil	.....	.....	A. Illite and Chlorite	Jeffries and Yearick, 1949
	Union Co., Penn.	.....	.....		
	Pleistocene glacial drift	.....	.....	C. Illite and Chlorite	

TABLE 1 (*Cont.*)

Chemical Data and Clay Mineral Content of Colloidal Fraction of Soils and Sediments Discussed in Text

A. A horizon of soil
B. B horizon of soil
C. Parent material of C horizon

I — Illite group
M — Montmorillonite group
K — Kaolinite group

Soil	%K_2O	molecular SiO_2/Al_2O_3	Predominant clay mineral	Reference (Tech. Bulls. cited are of U. S. Dept. Agr.)
27. Norfolk soil, Red and Yellow Podzolic soil	A. 0.52	A. 1.90	A. Kaolinite	Tech. Bull. 594
eastern North Carolina, Georgia, and Alabama	B. 0.53	B. 1.86		Pearson and Ensminger, 1949
Cenozoic deposits of Atlantic and Gulf coastal plains	C. 0.30	C. 1.93	C. Kaolinite	
28. Oktibbeha clay, Rendzina	.....	.....	A. M and K	Pearson and Ensminger, 1949
Black belt, central Ala.	.....	.....		
Upper Cretaceous marine deposits	.....	.....	C. M and K	
29. Paleosol, Pre-Wisconsin age	.....	A. 7.22	A. Illite	Hunt and Sokoloff, 1950
Provo Canyon, Utah	.....	.....		
Upper Paleozoic Manning Canyon shale (marine)	.....	.....	C. Illite?	
30. Palouse silt loam, Chernozem soil	A. 2.12	A. 3.23		Tech. Bull. 316
3 miles northwest of Pullman, Whitman Co., Wash.	B. 1.64	B. 3.24	B. K-bentonite type	
Pleistocene loess (non-marine)	C. 1.46	C. 3.50		
31. Pawnee silt loam, Prairie soil	.....	.....	A. Illite, in coarse clay fraction	Larson, *et al.*, 1947
Gage Co., Nebraska	.....	.....		
Kansan glacial till	.....	.....	C. Illite, in coarse clay fraction	
32. Pierre clay, Lithosol	A. 2.14	A. 3.76		Tech. Bull. 502
Gregory Co., South Dakota	B. 2.14	B. 3.93		
Upper Cretaceous Pierre shale (marine)	C. 2.15	C. 4.01		
33. Pullman loam, Reddish Chestnut soil	A. 2.68	A. 3.89		Tech. Bull. 502
Masterson, Nash Co., Texas	B. 2.50	B. 3.87		
Pliocene Ogallala formation (non-marine)	C. 2.26	C. 3.37		
34. Putnam silt loam, Planosol	A. 0.94	A. 3.86	A. Mixed layer aggregate or beidellite	Ross and Hendricks, 1945
northeastern Missouri	.....	.....		
Pleistocene loess (non-marine)	.....	.....	C. M and I?	
35. Scobey loam, Chestnut soil	A. 2.35	A. 4.47		Tech. Bull. 502
McKensie Co., North Dakota	B. 1.90	B. 3.97		
Wisconsin glacial drift (calcareous)	C. 1.80	C. 4.34		
36. Scott silt loam, Wiesenboden soil	.....	.....	A. Illite, in coarse clay fraction	Larson, *et al.*, 1947
Saunders Co., Nebraska	.....	.....		
Pleistocene loess (non-marine)	.....	.....	C. Illite, in coarse clay fraction	
37. Shelby silt loam, Planosol	A. 1.19	A. 3.31		Tech. Bull. 316
6 miles west of Bethany, Harrison Co., Mo.	B. 1.62	B. 3.57	B. K-bentonite type	
Kansan glacial drift	C. 1.66	C. 3.54		
38. Spearfish silt loam, Lithosol	A. 2.33	A. 3.35		Tech. Bull. 502
Piedmont, Mead Co., South Dakota	B. 4.07	B. 3.06		Van Houten, 1948
Triassic Spearfish formation	C. 4.65	C. 2.97	C. Illite	
39. Susquehanna soil, Red and Yellow Podzolic soil	A. 0.34	A. 2.73	A. Kaolinite	Kelley, *et al.*, 1939
23 miles southeast of Montgomery, Montgomery Co., Ala.	.....	.....		Pearson and Ensminger, 1949
Cretaceous deposits, Gulf Coastal Plain (marine)	C. 0.26	C. 3.95	C. Montmorillonite	
40. Vernon sandy loam, Reddish Prairie soil	A. 2.13	A. 3.45		Tech. Bull. 316
4 miles south of Guthrie, Logan Co., Okla.	B. 2.34	B. 2.94	B. K-bentonite type	
Permo-Triassic red beds	C. 2.41	C. 2.94		
41. White Store Soil, Red and Yellow Podzolic soil	.....	.....	A. Montmorillonite	Ross and Hendricks, 1945
eastern Triassic basin, N. Carolina	.....	.....		
Triassic shale and sandstone (non-marine)	.....	.....	C. Montmorillonite	
II Sediments and Alluvial Soils (little or no profile)				
42. Red and sandy silt; Little Colorado River, Arizona	2.44	4.45	I, M, and K in equal amounts	Grim, *et al.*, 1949
43. Gray silt; center Mission Bay, San Diego, Calif.	2.39	3.81	I, M, and K in equal amounts	Grim, *et al.*, 1949
44. Green mud; west of Gulf of California	2.97	4.81	40%±I, 30%±M, 20%±K	Grim, *et al.*, 1949
45. Green silty clay; about 6 miles west of Pacific coast, northwest of San Diego, Calif., 275 fms.	2.69	4.00	40%±I, 30%±M, 20%±K	Grim, *et al.*, 1949
46. Mud from continental slope, about 175 miles west of San Diego, Calif., 1717 fms.	2.50	5.78	40%±I, 30%±M, 20%±K	Grim, *et al.*, 1949
47. Huntington loam, Alluvial soil; 3 miles northeast of Goconda, Ill., in Livingston Co., Ky.	3.40	2.78	70%I, 15%K, 10%M	Tech. Bull. 883
48. Huntington silt loam, Alluvial soil; Smithland, Livingston Co., Ky.	2.81	2.90	70%I, 10%K, 10%M	Tech. Bull. 883
49. Sharkey silty clay loam, Alluvial soil; 6 miles south of Onward, Route 61, Sharkey Co., Miss.	2.74	3.79	50%I, 35%M, 10%K	Tech. Bull. 883
50. Wabash silt loam, Alluvial soil; New Albin, Allamakee Co., Iowa	1.51	4.32	50%I, 35%M, 10%K	Tech. Bull. 883
51. Harve silty clay loam, Alluvial soil; Yellowstone River Bridge, Richland Co., Mont.	2.64	3.67	50%I, 35%M, 15%K	Tech. Bull. 883
52. Miller clay, Alluvial soil; Alexandria, Rapides Parish, La.	3.15	3.70	60%I, 20%M, 15%K	Tech. Bull. 883
53. Bibb silt loam, Alluvial soil; 9 miles southwest of Bastrop, Morehouse Parish, La.	1.85	3.22	70%I, 15%K, 10%M	Tech. Bull. 883
54. Sediment, Mississippi River mouth, Burrwood, Plaquemines Parish, La.	2.44	4.08		Tech. Bull. 883
55. Sediment, 3 miles southeast of Mississippi Delta, 70 fms.	2.62	3.78		Tech. Bull. 883
56. Sediment, Gulf of Mexico, two-thirds distance from delta to Florida, 1770 fms.	2.48	4.42		Tech. Bull. 883

tionation of colloidal material through the profile, or that clay minerals inherited from parent rock may be altered by appropriate weathering conditions (see Jackson, *et al.*, 1948). Furthermore, a soil colloid containing some inherited clay may be composed largely of primary clay minerals produced by weathering of detrital minerals of the parent rock. It is also recognized that the coarser and finer colloidal fractions generally contain different proportions of the clay minerals (Bray, 1934; Keller and Ting, 1950), and that few soil clays consist of a single clay mineral. Instead, most contain a complex mixture of very small units of several clay minerals, although one mineral commonly is dominant. In this discussion illite (illitic), montmorillonite (montmorillonoid), and kaolinite (kaolinic) refer to groups of clay minerals, not to particular minerals.

Intense weathering has converted montmorillonite in the Cenozoic deposits of the Alabama Coastal Plain to kaolinite in the overlying Susquehanna soil (39). The difference in clay content of the soil and parent material is clearly reflected in the molecular SiO_2/Al_2O_3 ratios of the A (2.73) and C (3.95) horizons—a 1.22 difference. In contrast to this example of an authigenic soil clay, the kaolinite-rich Decatur soil (17) and Norfolk soil (27) and the montmorillonite-rich Houston and related soils (18, 19, 9, 28) undoubtedly inherited their clay minerals from parent material. In these soils the same clay minerals predominate in the A and C horizons and there are similar SiO_2/Al_2O_3 ratios throughout each profile.

In many soils developed on sedimentary parent material illite is abundant in the A and C horizons (table 1; 6, 8, 12, 25, 26, 31, 36), or there are similar SiO_2/Al_2O_3 ratios throughout each profile (no difference exceeds 0.51) and illite has been identified as the dominant clay mineral[3] in the B horizons (table 1; 2, 3, 4, 5, 16, 17, 22, 24, 30, 37, 40). These data reveal that in many different kinds of soil widely distributed throughout the United States (fig. 1) illite is the abundant clay mineral and that much of it undoubtedly was derived from illite-rich parent material. Further suggestive evidence is provided by an illite-bearing Paleosol (29) in the Rocky Mountain region that developed on Paleozoic marine shale. Because illite is known to be abundant

[3] K-bentonite type clay mineral is considered to be an illite mineral (Weaver and Bates, 1950) or a mixed layer aggregate with layers of illite (Kerr, *et al.*, 1950*a* and *b*).

in such shale, it is reasonable to conclude that the illite in the Paleosol may have been inherited from an illite-rich parent rock.

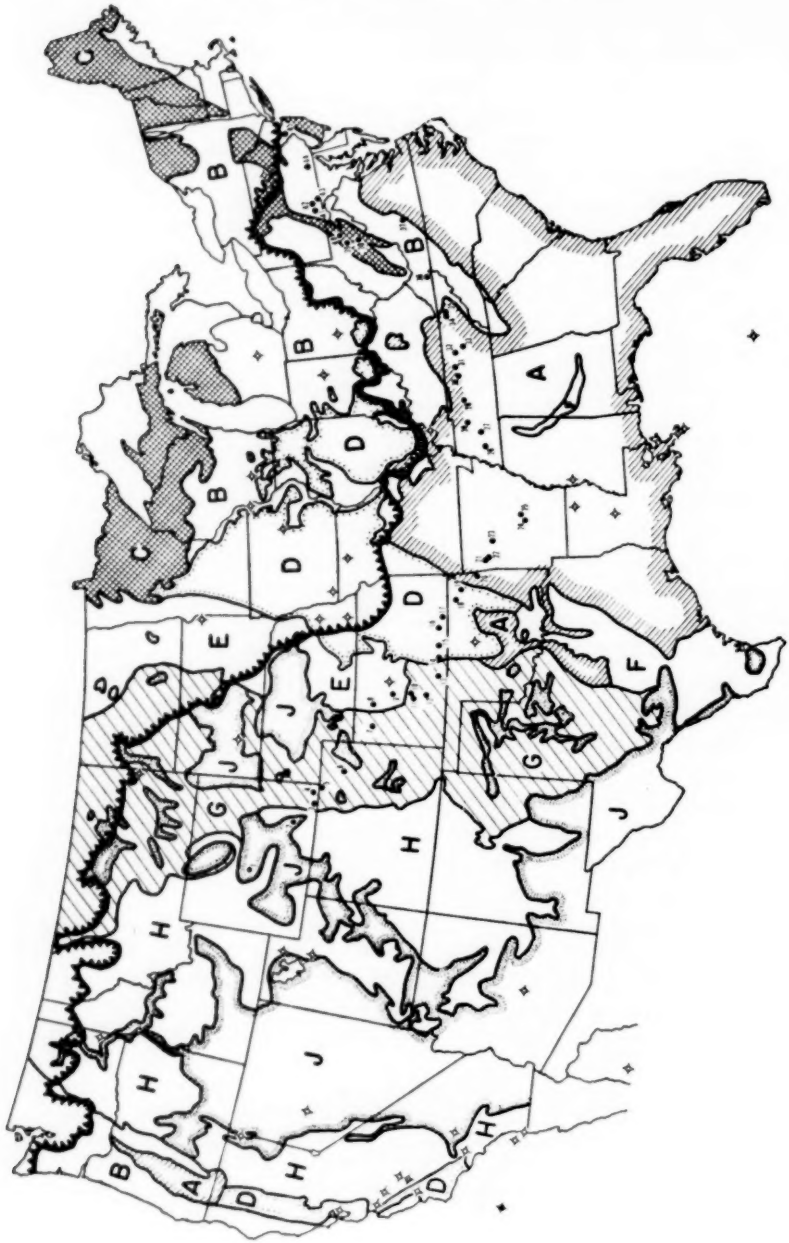
In addition to the persistently high SiO_2/Al_2O_3 ratios in the profiles of illite-bearing soils, the K_2O content of their soil colloids is, with but few exceptions, higher than in soil colloids rich in kaolinite and montmorillonite minerals.[4] Moreover, the K_2O content of many of these soils is remarkably like that of illite-rich sediments in the Pacific Ocean and Gulf of California (table 1; 43-46; also fig. 3; A, B, E, F). The amount commonly present (1 to 4 per cent) is, however, less than in Fithian[5] illite (table 3; d). In fact, the illite-rich Cayucos soil colloid (6) contains as little as 0.66 per cent K_2O, and Ross and Hendricks (1945, p. 61) have concluded that the presence of 0.94 per cent K_2O in the Putnam soil (34) reveals that its clay mineral may be a mixed layer aggregate with layers of illite.[6] Apparently more than 1 per cent K_2O in the colloidal fraction is suggestive evidence of the presence of some illite or illite layers in a mixed layer mineral. Hendricks and Fry (1930) first pointed out that the kind of illite in soil colloids contains less K_2O than the illite in Ordovician K-bentonite. More recent work has shown that the illite mineral abundant in soils is a degraded form produced during weathering by the leaching of some K^+ from between the silicate layers (Grim, 1951). The fact that in some of the illite-bearing profiles recorded in table 1 (3, 4, 17, 24) the K_2O content of the A horizon colloidal fraction is appreciably less than that of the parent material may be evidence of the process in operation. The presence of a degraded form of illite in soil must be recognized in attempts to estimate the illite content from chemical data. Several published estimates have been made assuming a 4 to 6 per cent K_2O content, and as a result the proportion of illite reported probably is too small.

The kind of clay minerals present in the colloidal fractions of the Keith (20), Pullman (33), Scobey (35), and Spearfish (38) soil profiles have not been reported in the literature referred to. The molecular SiO_2/Al_2O_3 ratios and K_2O contents

[4] In early studies of soils a high potassium content of the colloids was attributed to the presence of unweathered feldspar.

[5] Illite from Pennsylvanian underclay at Fithian, Illinois, is a representative example of illite as defined by Grim, Bray, and Bradley, 1937.

[6] The whole colloid of the Putnam soil of Illinois contains 1.74 per cent K_2O, according to Bray, 1934.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show little variation throughout the profiles and are similar to those of many illite-rich soil colloids. Moreover, illite is the predominant clay mineral in analyzed samples of the Spearfish formation (Van Houten, 1948, p. 2100) and the other soils are developed on sedimentary materials in which illite commonly is abundant. The evidence strongly suggests that these soils may contain inherited illite. But chemical composition generally is not a reliable guide for the prediction of the clay minerals present, especially since the colloid may contain some uncombined silica or alumina and a montmorillonite clay mineral with adsorbed K^+ may have a composition much like degraded illite. In fact, the SiO_2/Al_2O_3 ratios of these profiles are not unlike that of montmorillonite (table 3; f and g), and this clay mineral has been reported (Ross and Hendricks, 1945, p. 63) to be a prominent constituent of northern Prairie and Chernozem soils developed on calcareous glacial drift and loess (cf. Keith and Scobey soils). However, Ross and Hendricks do suggest that the clay in these soils is partly inherited from parent material. Montmorillonite may also be the predominant clay mineral in the Pierre shale (Grim, 1951, p. 228) parent rock of the Pierre soil (32).

A number of reported montmorillonite-rich soils are based on the identification of beidellite in colloidal fractions containing more than 1 per cent K_2O. It is significant, therefore, that some of the clay minerals that have been called beidellite actually are a mixture in which illite is an important component

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Fig. 1. Generalized map of great soil groups of United States.

- A. Red and Yellow Podzolic soils
- B. Gray-Brown Podzolic soils
- C. Podzolic soils
- D. Prairie and Reddish Prairie soils, and Planosols
- E. Chernozem soils
- F. Rendzina soils
- G. Brown, Reddish Brown, Chestnut, and Reddish Chestnut soils
- H. Undifferentiated mountain soils, and complex areas of soils in California
- J. Gray Desert and Red Desert soils, and Lithosols

Barbed line, southern limit of Pleistocene glacial drift sheets.

Circle with cross, illite-rich soils and sediments recorded in literature (table 1).

Solid circle, soils and parent rock analyzed in this investigation (table 2).

(Grim and Rowland, 1942, p. 801-804; Grim, Dietz, and Bradley, 1949, p. 1788). Loess, for example, has been described as containing abundant beidellite, but at least some of it is composed predominantly of illite (Hursh, Lamar, and Grim, 1944, p. 22) and the colloidal fraction of loess-derived soils of Illinois contains a mixture of illite and montmorillonite (beidellite) (Grim, 1941, p. 13; Kurtz, DeTurk, and Bray, 1946, p. 114). The Putnam soil of northeastern Missouri (table 3; k) has been reported as beidellite-rich, but it may, instead, contain a mixed layer mineral with layers of illite. Similarly, Kerr, *et al.* (1950*a*, p. 39) have shown that several clays reported to be beidellite are illite and montmorillonite mixed layer aggregates. In fact, in pointing out the common occurrence of mixed layer minerals, Hendricks and Alexander (1939, p. 261) expressed doubt that individual montmorillonite crystals are even common constituents of soils and concluded that much of the clay fraction of soils and sediments apparently is an intricate mixture of several clay minerals. Identification of these clay minerals is made still more difficult by the fact that some minerals of the montmorillonite and illite groups may have very similar physical properties (Weaver and Bates, 1951).

An analysis of alluvial[7] soils of the Mississippi drainage basin by Holmes and Hearn (1942) has revealed that the colloidal fractions of eastern tributary soils have consistently lower molecular SiO_2/Al_2O_3 ratios than do the colloidal fractions of western tributary alluvial soils. But the eastern alluvial soils have the highest average illite and K_2O content. The data in table 1 show similarly that the colloidal fractions of the uniform zonal[8] soil profiles in areas drained by eastern tributaries (table 1; 3, 4, 17, 24) have lower SiO_2/Al_2O_3 ratios than those of soil profiles in areas drained by western Mississippi tributaries (table 1; 2, 5, 16, 22, 37, 40; table 2; 13, 14, 17). Comparative data are assembled in table 1a.

Because both zonal and alluvial soils in the east have low SiO_2/Al_2O_3 ratios and yet contain relatively as much or more illite than their western counterparts the low ratio cannot be attributed to the presence of kaolinite. These data suggest,

[7] Alluvial soils are composed of transported soil material on which little or no profile has developed.

[8] Zonal soils have well-developed profiles that reflect the influence of the active factors of soil genesis.

instead, that there is uncombined alumina in the eastern colloids or that weathering processes in the more humid east may have removed some silica from or introduced alumina into the illite present throughout the profile. Moreover, in the east the A horizon illite apparently is more degraded with respect to K_2O than the illite in the lower part of the profile, for the eastern zonal soils have a consistently lower K_2O content in

TABLE 1a

Average Compositions

	Molecular SiO_2/Al_2O_3 ratio	per cent K_2O	Clay Minerals Present
A. Alluvial Soils			
Eastern tributaries	2.86	3.05	70% Illite, 10-15% Kaolinite
Western tributaries	3.87	2.58	50-60% Illite, 10-15% Kaolinite
B. Zonal Soils			
Eastern tributaries	A. hor. 3.32	A. hor. 2.08 C. hor. 3.78	65-75% Illite, 10% Kaolinite in B. hor.
Western tributaries	A. hor. 4.14	A. hor. 2.28 C. hor. 2.55	Abundant Illite in B. hor.

their A horizons, whereas in western soils there commonly is no appreciable difference in K_2O content throughout the profiles. If these samples of the zonal soils are representative, it is not clear why eastern alluvial soils have a higher K_2O content than the western ones, inasmuch as alluvial soils are presumably composed largely of little-altered mineral eroded from the higher horizons of zonal soil profiles. The evidence implies, instead, that the C horizon must have contributed much material to the eastern alluvial soil or that there is no direct correlation between the composition of zonal and alluvial soils of the same region.

B

In order to provide more extensive information about the relationship of clay minerals in soils and their parent sedimentary rocks, samples of sedimentary rocks and the A horizon of derived soils from many soil areas (table 2) have been analyzed for the predominant clay mineral group present. The samples were collected south of the southern limit of the glacial

drift sheets, along a route from Wyoming to Oklahoma and Arkansas to New Jersey (fig. 1).

The colloidal fractions of the pairs of samples were treated with H_2O_2 and HCl, and the more abundant clay minerals present were identified by differential thermal and x-ray analysis. In many of the samples the subordinate clay constituents have not been determined, hence these specimens may contain some clay minerals not reported in table 2. Where the presence of montmorillonite was suggested by the differential thermal curves or by published analyses of similar soils, those clays were x-rayed both untreated and after being soaked in glycol. In the clay fractions identified here as "illite and montmorillonite," evidence of these minerals was found in both differential thermal and x-ray curves. Clay fractions recorded as "K-bentonite type" yield a differential thermal curve like that of mixed illite and montmorillonite but an x-ray curve of illite. Some authors consider that K-bentonite is a type of illite, others believe it is a mixed layer mineral or aggregate composed of illite and montmorillonite. The chemical analyses recorded in table 3 (a, b, c) were made by the Rock Analysis Laboratory, Department of Geology, University of Minnesota.

In 35 of the 43 pairs of samples with sedimentary bedrock the predominant clay mineral groups in the soil are the same as those in the parent rock member (table 2). Nevertheless, more detailed analysis undoubtedly would show differences in the proportions of the clay minerals present. It is to be expected that inherited clay minerals characterize the immature lithosols of the Appalachian Mountains (table 2; 32, 33, 40, 41, 42; fig. 2; 33, 42) and the shallow Sierozem (table 2; 1, 2; fig. 2; 1) and Brown (table 2; 3, 6) soils of arid regions. The significant fact revealed by these analyses is the similarity of the predominant clay minerals in most of the samples of zonal soils and their parent rocks from humid regions. Even in southern United States where intense weathering conditions commonly produce primary clay minerals in the soils, many of the kaolinite-rich Red and Yellow Podzolic soils (table 2; 18, 19, 21-27; fig. 2; 25, 27) apparently have inherited their clay from kaolinite-rich parent material, as did the Norfolk soils (table 1; 27).

Of the eight soils whose clay mineral content differs markedly from that of the parent material, one is a kaolinite-rich soil developed on the Boone chert (table 2; 20). The other seven

soils also contain abundant kaolinite, but they have been formed on limestone, dolomite, and calcareous shale rich in illite or a K-bentonite type of clay mineral (table 2; 28, 29, 30, 31, 35, 36, 43). Numerous published analyses show that Paleozoic limestone and dolomite, with a few kaolinite-rich exceptions, contain abundant illite (Grim, Lamar, and Bradley, 1937; Grim, 1951, p. 227), and that soils developed on the Paleozoic calcareous rocks of the Appalachian region (table 1; 7, 10, 11, 14, 15, 23) are consistently kaolinite-rich with some illite (Alexander, Byers, and Edgington, 1939). Apparently the paucity of clay in the parent rock and the acid environment of Red and Yellow Podzolic soils favor the development of abundant primary kaolinite. Nevertheless, some of the predominant illite of the parent rock does persist as inherited illite in the derived soils.

Montmorillonite is an important constituent in many of the analyzed soils and rocks. No attempt has been made to estimate the relative amounts of illite and montmorillonite in most samples rich in these minerals, although the differential thermal curves do suggest that illite is consistently abundant. Significantly, most of the soils rich in montmorillonite (table 2; 1-16) are developed on montmorillonite-rich parent rock. In fact, some of these parent rocks contain relatively more montmorillonite than their derived soils (table 2; 1, 2, 3, 7). This association reveals that at least part of the montmorillonite so prevalent in soils of the western plains has been inherited from montmorillonite-bearing sediments. Nevertheless, montmorillonite in some of the analyzed soils (table 2; 9, 11) is clearly primary, for here it is relatively more abundant than in the parent rocks.

The results of these analyses also reinforce the conclusion that illite is a very common clay mineral group in sedimentary rocks and that it is a "relic" mineral in many soils. The significant resemblance of the differential thermal curves of illite-bearing zonal soils and parent materials (fig. 2; 13, 14, 16) is emphasized by comparison with curves of kaolinite-rich Gray-Brown Podzolic soils developed on mica schist (fig. 2; 37, 44). The presence of illite in these samples does not mean, however, that the colloid contains a clay mineral with as high a K_2O content and molecular SiO_2/Al_2O_3 ratio as Fithian illite. Chemical analyses of the colloidal fraction of three pairs of illite-bearing samples (table 2; 13, 14, 17; table 3; a, b, c) show that their K_2O content and SiO_2/Al_2O_3 ratios are within the range of

TABLE 2

Soils and Parent Sedimentary Rocks from Numbered Localities on Figure 1.

A. Soil, A horizon
C. Parent rock

I — Illite group
M — Montmorillonite group
K — Kaolinite group

	Soil	Predominant clay mineral group
1.	A. Navajo-Chipeta soils, Sierozem soil group (fig. 2; see table 1;1)	Illite, some Montmorillonite
	25 miles southeast of Riverton, Fremont Co., Wyo.	
	C. Oligocene White River tuffaceous sandstone (non-marine)	Illite and Montmorillonite
2.	A. Navajo-Chipeta soils, Sierozem soil group	Illite and Montmorillonite
	Route 30, 2 miles east of Rock River, Albany Co., Wyo.	
	C. Upper Cretaceous Steele shale and sandstone (marine)	Montmorillonite and Illite
3.	A. Joplin soil, Brown soil group	Illite, some Montmorillonite
	Route 30, 11 miles west of Cheyenne, Laramie Co., Wyo.	
	C. Pliocene Ogallala arkose (non-marine)	Illite and Montmorillonite
4.	A. Rosebud-Bridgeport soils, Chestnut soil group	Illite and Montmorillonite
	Route 30, 20 miles east of Cheyenne, Laramie Co., Wyo.	
	C. Pliocene Ogallala arkosic conglomerate (non-marine)	Illite and Montmorillonite
5.	A. Rosebud-Bridgeport soils, Chestnut soil group (see table 1;1)	Illite and Montmorillonite
	Route 155, 2 miles south of Hereford, Weld Co., Colo.	
	C. Oligocene White River tuffaceous siltstone (non-marine)	Illite and Montmorillonite
6.	A. Fort Collins soils, Brown soil group	Illite and Montmorillonite
	Route 34, 2 miles west of Akron, Washington Co., Colo.	
	C. Pliocene Ogallala arkose (non-marine)	Illite and Montmorillonite
7.	A. Keith soil, Chestnut soil group (fig. 2; see table 1; 20)	Illite and Montmorillonite, little K
	Route 24, 16 miles west of Colby, Thomas Co., Kan.	
	C. Pliocene Ogallala arkosic sandstone (non-marine)	Illite and Montmorillonite, little K
8.	A. Holdrege-Hall soils, Chernozem soil group	Illite and Montmorillonite
	Route 24, 1.5 miles west of Hoxie, Sheridan Co., Kan.	
	C. Pliocene Ogallala arkosic sandstone (non-marine)	Illite and Montmorillonite
9.	A. Hays soil, Chernozem soil group (see table 1;16)	Illite and Montmorillonite
	Route 283, 1 mile south of Smoky Hill River, Trego Co., Kan.	
	C. Upper Cretaceous Niobrara chalk (marine)	Illite, little Montmorillonite
10.	A. Holdrege-Hall soils, Chernozem soil group (fig. 2)	Illite and Montmorillonite
	Route 283, 1 mile north of Jetmore, Hodgeman Co., Kan.	
	C. Pleistocene loess (non-marine)	Illite and Montmorillonite
11.	A. Hays soil, Chernozem soil group (see table 1;16)	Illite and Montmorillonite
	Route 50, 5 miles west of Jetmore, Hodgeman Co., Kan.	
	C. Upper Cretaceous Carlisle calcareous shale (marine)	Illite, some Montmorillonite
12.	A. Holdrege-Hall soils, Chernozem soil group	Illite and Montmorillonite
	14 miles west of Ford, Ford Co., Kan.	
	C. Pleistocene loess (non-marine)	Illite and Montmorillonite
13.	A. Miles-Vernon soils, Reddish Chestnut soil group (fig. 2; table 3,a; see table 1;40)	Illite and Montmorillonite
	Route 160, 15 miles west of Medicine Lodge, Barber Co., Kan.	
	C. Permian Cimarron red siltstone	Illite and Montmorillonite
14.	A. Zaneis-Renfrow soils, Reddish Prairie soil group (fig. 2; table 3,b)	Illite, some Montmorillonite
	Route 160, 7 miles east of Attica, Harper Co., Kan.	
	C. Permian Cimarron red siltstone	Illite, some Montmorillonite
15.	A. Zaneis-Renfrow soils, Reddish Prairie soil group	Illite, some Montmorillonite
	Route 160, 17 miles west of Wellington, Summer Co., Kan.	
	C. Permian Cimarron sandstone	Illite, some Montmorillonite
16.	A. Summit-Bates soils, Prairie soil group (fig. 2)	Illite, some Montmorillonite
	Route 160, 4 miles east of Winfield, Cowley Co., Kan.	
	C. Permian Big Blue calcareous shale (marine)	Illite, some Montmorillonite
17.	A. Florence-Newtonia soils, Prairie soil group (see table 3;c)	Illite and Kaolinite
	Route 166, 1 mile east of Cedarvale, Chautauqua Co., Kan.	
	C. Pennsylvanian Wabaunsee sandstone (cyclothem)	Illite and Kaolinite
18.	A. Parsons and associated soils, Planosol group	Kaolinite and Illite
	Route 60, 6 miles east of Nowata, Nowata Co., Okla.	
	C. Pennsylvanian Coffeyville micaceous siltstone	Kaolinite, some M. and Illite
19.	A. Parsons and associated soils, Planosol group	Kaolinite, some Illite
	Route 67, 17 miles north of Pryor, Craig Co., Okla.	
	C. Pennsylvanian Cherokee sandstone (cyclothem)	Kaolinite, some Illite and M.
20.	A. Baxter-Lebanon soils, Red and Yellow Podzolic soil group (see table 1;21)	Kaolinite
	Route 33, 4 miles west of Kansas, Delaware Co., Okla.	
	C. Mississippian Boone chert and cherty limestone (marine)	Chert
21.	A. Hanceville-Conway soils, Red and Yellow Podzolic soil group	Kaolinite, some Illite
	Route 71, 28 miles south of Fayetteville, in Crawford Co., Ark.	
	C. Pennsylvanian Atoka gray shale	Kaolinite, some Illite
22.	A. Hanceville-Conway soils, Red and Yellow Podzolic soil group	Kaolinite, some Illite
	Route 64, 9 miles west of Ozark, Franklin Co., Ark.	
	C. Pennsylvanian shale	Kaolinite, some Illite and M.
23.	A. Hanceville-Conway soils, Red and Yellow Podzolic soil group	Kaolinite and Illite
	Route 64, 7 miles east of Clarksville, Johnson Co., Ark.	
	C. Pennsylvanian McAlester shale	Kaolinite, little Illite

TABLE 2 (*Cont.*)

Soils and Parent Sedimentary Rocks from Numbered Localities on Figure 1.

A. Soil, A horizon
C. Parent rock

I — Illite group
M — Montmorillonite group
K — Kaolinite group

	Soil	Predominant clay mineral group
24.	A. Norfolk-Ruston soils, Red and Yellow Podzolic soil group (See table 1;27)	Kaolinite
	Route 270, 1 mile west of Sheridan, Grant Co., Ark.	
	C. Eocene Claiborne sandy clay (lagoonal)	Kaolinite
25.	A. Caddo-Beauregard soils, Red and Yellow Podzolic soil group (fig. 2)	Kaolinite
	Route 167, 19 miles north of Sheridan, Grant Co., Ark.	
	C. Eocene Claiborne sandstone (lagoonal)	Kaolinite
26.	A. Memphis-Grenada soils, Red and Yellow Podzolic soil group	Kaolinite
	Route 79, 6 miles east of Stanton, Haywood Co., Tenn.	
	C. Eocene Grenada sandy clay (Wilcox group) (lagoonal)	Kaolinite
27.	A. Memphis-Grenada soils, Red and Yellow Podzolic soil group (fig. 2)	Kaolinite, some Illite
	Route 70, 4 miles northeast of Jackson, Madison Co., Tenn.	
	C. Eocene Wilcox sandy clay (lagoonal)	Kaolinite
28.	A. Dickson-Baxter soils, Red and Yellow Podzolic soil group	Kaolinite and Illite
	Route 70, 1 mile east of Camden, Benton Co., Tenn.	
	C. Devonian calcareous shale and limestone (marine)	K-bentonite type, a little Kaolinite
29.	A. Dickson-Baxter soils, Red and Yellow Podzolic soil group	Kaolinite, some Illite
	Route 70, 2 miles east of Dickson, Dickson Co., Tenn.	
	C. Mississippian limestone (marine)	Illite and Montmorillonite, little K.
30.	A. Maury soil, Red and Yellow Podzolic soil group, (fig. 2; see table 1; 23)	Kaolinite
	Route 70N, 8 miles east of Lebanon, Wilson Co., Tenn.	
	C. Ordovician phosphatic limestone (marine)	Illite
31.	A. Maury soil, Red and Yellow Podzolic soil group (see table 1, 23)	Illite and Kaolinite
	Route 70N, 9 miles east of Carthage, Smith Co., Tenn.	
	C. Ordovician phosphatic limestone and shale (marine)	Illite
32.	A. Hartsells-Muskingum soils, Lithosol group (Red and Yellow Podzolic soil group)	Kaolinite and Illite
	Route 70N, 3 miles west of Monterey, Putnam Co., Tenn.	
	C. Mississippian sandstone and shale	Kaolinite, some Illite
33.	A. Hartsells-Muskingum soils, Lithosol group (Red and Yellow Podzolic soil group)	Illite and Kaolinite
	Route 70, 3 miles east of Ozone, Cumberland Co., Tenn.	
	C. Pennsylvanian Lee sandstone and shale (delta, non-marine?)	Illite and Kaolinite
34.	A. Decatur-Dewey-Clarksville soils, Red and Yellow Podzolic soil group	Illite, some Kaolinite
	Route 25E, 9 miles west of Mooresburg, Grainger Co., Tenn.	
	C. Mississippian carbonaceous shale (non-marine)	Illite
35.	A. Hagerstown-Frederick soils, Gray-Brown Podzolic soil group (see table 1;10,14,15)	Kaolinite and Illite
	Route 11W, 3 miles northeast of Mooresburg, in Hawkins Co., Tenn.	
	C. Ordovician Calcareous reddish brown shale (marine)	Illite (K-bentonite type?)
36.	A. Hagerstown-Frederick soils, Gray-Brown Podzolic soil group (fig. 2; see table 1;10,14,15)	Kaolinite and Illite
	Route 19, 1 mile west of Abingdon, Washington Co., Va.	
	C. Ordovician Knox dolomite (marine)	K-bentonite type
37.	A. Porters-Ashe soils, Gray-Brown Podzolic soil group (fig. 2)	Kaolinite
	2 miles east of Lynchburg, Campbell Co., Va.	
	C. Precambrian mica schist	Micaceous material
38.	A. Dekalb-Leetonia soils, Podzol soil group	Illite, some Kaolinite
	Route 250, 15 miles south of Elkins, Randolph Co., W. Va.	
	C. Upper Devonian shale (marine)	Illite, M., some Kaolinite
39.	A. Muskingum-Wellston-Zanesville soils, Gray-Brown Podzolic soil group	Kaolinite and Illite
	Route 250, 3 miles northwest of Phillipi, Barbour Co., W. Va.	
	C. Pennsylvanian Conemaugh sandstone and shale (non-marine)	Kaolinite and Illite
40.	A. Muskingum soil, Lithosol group (Podzol soil group)	Illite
	Route 50, 4 miles east of Aurora, Preston Co., W. Va.	
	C. Mississippian Pocono gray sandstone and shale (non-marine)	Illite
41.	A. Muskingum soil, Lithosol group (Gray-Brown Podzolic soil group)	Illite
	Penn. Turnpike, Clear Ridge Cut, Bedford Co., Penn.	
	C. Upper Devonian Marcellus shale (marine)	Illite
42.	A. Muskingum soil, Lithosol group (Gray-Brown Podzolic soil group) (fig. 2)	Illite and Kaolinite
	Penn. Turnpike, Tuscarora Tunnel, Huntingdon Co., Penn.	
	C. Upper Devonian Portage shale (marine)	Illite and Kaolinite
43.	A. Hagerstown soil, Gray-Brown Podzolic soil group (fig. 2; see table 1;14,15)	Illite and Kaolinite
	Route 11, 8 miles south of Chambersburg, Franklin Co., Penn.	
	C. Lower Ordovician limestone (marine)	K-bentonite type
44.	A. Manor soil, Gray-Brown Podzolic soil group (fig. 2)	Kaolinite
	Safe Harbor, Lancaster Co., Penn.	
	C. Precambrian? Wissahickon schist	Micaceous material
45.	A. Penn soil, Gray-Brown Podzolic soil group	Illite
	5 miles north of Princeton, Mercer Co., New Jersey	
	C. Triassic Brunswick red-brown shale (non-marine)	Illite

composition of the illite-bearing soils listed in table 1. The MgO content of no. 13 and 14 soil and "red bed" parent rock is unusually high, however. The Spearfish soil (table 1; 38) with 10.34 per cent MgO in the C horizon parent material (Spearfish "red bed" formation) is the only one listed with a colloidal fraction MgO content exceeding 4 per cent. The fact that each of these colloids has a low CO_2 content suggests that much of the magnesium is associated with a colloidal silicate mineral. The presence of a magnesium-rich colloid in the Per-

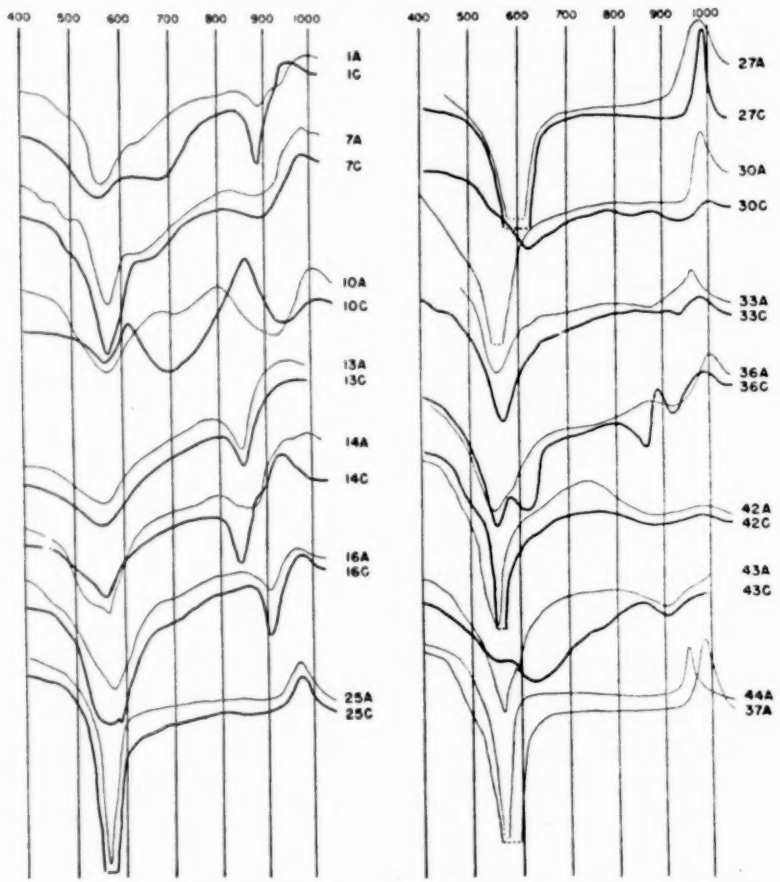

Fig. 2. Differential thermal curves of clay fraction of some rocks and soils listed in table 2.

mian and Triassic "red beds" may possibly be related to the environmental condition under which they formed. If so, it is perhaps significant that the magnesium-rich clay minerals such as attapulgite and sepiolite are abundant in highly alkaline sediments deposited in dry desert basins (Grim, 1951, pp. 228-229).

DISCUSSION OF SOME SOILS AND PARENT ROCKS LISTED IN TABLE 2:

1. The White River formation in this area contains both volcanic ash and sticky "bentonitic" clay. X-ray and differential thermal curves indicate that montmorillonite is more abundant in the parent rock than it is in the derived soil.

2 and 3. Both x-ray and differential thermal curves indicate that montmorillonite is relatively more abundant in the rocks than it is in the soils.

5. The White River formation in Colorado contains volcanic ash as well as some montmorillonite, but apparently not as much of this type of clay as is present in rock sample no. 1. The chemical composition of the rock probably is very similar to that of the Badland lithosol from western South Dakota (table 1; 1) which is developed on the White River formation. The presence of 2.61-2.75 per cent K_2O in the published profile is undoubtedly due to the presence of considerable illite.

3, 4, 6, 7, and 8. The Ogallala formation contains volcanic ash and montmorillonite clay which, according to x-ray analyses, is most abundant in rock sample no. 7. But no great abundance of montmorillonite is indicated by the differential thermal curve (fig. 2).

9 and 11. The similarity of clay content of soil and parent rock accords with the uniform chemical composition of the Hays profile in table 1 (16) which is also developed on Carlisle calcareous shale. The fact that illite appears to be the more abundant clay mineral present agrees with the reported abundance of K-bentonite type mineral in Hays soil (Middleton, Slater, and Byers, 1932).

10 and 12. The analyses of the parent material agree with published reports that loess contains illite and montmorillonite.

13. The Vernon soil colloid in table 1 (40) has a similar K_2O content and is similarly rich in illite for it has been reported to be composed of a K-bentonite type. The higher K_2O content of the parent rock of sample no. 13 suggests that there may

TABLE 3
Chemical Analysis

	a		b		c		d	e	f	g	h	i	j	k
	C	A	C	A	C	A								
SiO_2	50.66	39.59	47.49	50.80	52.04	50.79	56.91	57.41	50.95	57.49	44.39	67.51	53.43	49.50
Al_2O_3	16.29	15.46	18.17	17.84	24.73	18.61	18.50	17.96	16.54	20.27	18.11	15.09	21.59	21.75
Fe_2O_3	6.30	5.49	8.82	7.36	4.04	6.45	4.99	4.99	1.36	2.92	1.60	4.29	1.04	8.81
FeO	0.89	0.61	0.84	0.73	1.18	0.81?	0.26	0.26	0.26	0.19	0.50	1.26	tr	0.42
MgO	7.47	8.73	6.92	4.12	1.59	1.55	2.07	2.25	4.65	3.18	2.24	1.55	4.17	1.99
CaO	1.36	6.91	1.26	2.00	0.70	3.00	1.59	0.64	2.26	0.23	11.35	0.11	0.53	0.28
Na_2O	0.52	0.24	0.60	0.64	0.58	0.29	0.43	0.15	0.17	1.32	0.64	0.40	0.11	0.44
K_2O	3.17	2.32	4.37	4.09	4.34	2.87	5.10	5.75	0.47	0.28	4.40	2.77	7.00	0.94
H_2O+	6.37	7.51	6.23	6.26	6.65	7.06	5.98	6.70	8.28	6.85	4.34	4.43	6.67	13.05
H_2O-	5.94	7.76	4.21	4.20	2.76	4.90	2.86	2.97	15.01	7.63	3.28	1.36	5.70	
CO_2	0.04	3.84	0.41	0.24	0.00	0.15	....	....	....	....	8.43	0.97	0.28	....
TiO_2	0.74	0.64	0.80	0.86	1.25	0.85	0.81	0.82	0.32	0.12	0.36	0.09	....	0.70
P_2O_5	0.04	0.26	0.08	0.23	0.18	0.33	....	....	....	....	0.11	....	....	2.16
MnO	0.05	0.05	0.05	0.08	0.02	0.06	....	....	.01	....	0.02	0.04	....	0.02
ign. loss	0.21	0.83	0.00	0.50	0.00	2.25	....	....	....	....	....	....	....	....
mol. SiO_2/Al_2O_3	5.30	4.35	4.44	4.82	3.59	4.62	5.23	5.42	5.18	4.39	4.15	7.56	4.45	3.86

a. see table 2; 13.
   A — soil, A horizon, colloidal fraction
   C — parent rock, colloidal fraction

b. see table 2; 14.
   A — soil, A horizon, colloidal fraction
   C — parent rock, colloidal fraction

c. see table 2; 17.
   A — soil, A horizon, colloidal fraction
   C — parent rock, colloidal fraction

d. Illite; Fithian, Ill., Kerr, *et al.*, 1950*b*.

e. Illite; Morris, Ill., Kerr, *et al.*, 1950*b*

f. Montmorillonite; Polkville, Miss., Kerr, *et al.*, 1950*b*

g. Montmorillonite; Upton, Wyo., Kerr, *et al.*, 1950*b*

h. Mixed layer aggregate (K-bentonite); Tazewell, Va., Kerr, *et al.*, 1950*b*

i. Mixed layer aggregate (K-bentonite); Catawba, Va., Kerr, *et al.*, 1950*b*

j. Mixed layer aggregate (K-bentonite); High Bridge, Ky., Kerr, *et al.*, 1950*b*

k. Putnam soil, Mixed layer aggregate or beidellite; Ross and Hendricks, 1945.

be more illite in the rock than in the derived soil. Although x-ray curves clearly indicate the presence of some montmorillonite in both soil and parent material, illite probably is the predominant clay mineral present as suggested by the differential thermal curves (fig. 2).

14 and 15. Differential thermal curves and the high K_2O content of sample no. 14 indicate an abundance of illite in both soil and rock. X-ray curves show that there is relatively less montmorillonite than in rock and soil no. 13.

18 and 19. The presence of some montmorillonite in the rocks and its absence in the derived soils strongly suggests that this clay mineral is actually destroyed by soil-forming processes here as in soils nos. 1, 2 and 3.

22 and 23. Both soils and parent rocks contain kaolinite and illite, but the soil of sample no. 23 contains relatively more illite and less kaolinite than its parent rock or than the same kind of soil of sample no. 22.

24. Clay mineral content is like that of the Norfolk soil and parent material in table 1 (27) and presumably the chemical compositions are similar.

27. The presence of some illite in the soil of a kaolinite-rich sample suggests that the illite may be primary, but this seems anomalous in a warm, humid climate where intense weathering readily produces kaolinite. A similar relative increase of illite over kaolinite was noted in sample no. 22.

30 and 31. The chemical composition probably is like that of Maury soil and parent material in table 1 (23). The higher K_2O content and molecular SiO_2/Al_2O_3 ratio in the C horizon correlates with the abundance of illite in the parent rock and its alteration to kaolinite in the soil.

33. The small endothermic deflections of about 630° in the differential thermal curves (fig. 2) of both rock and soil suggest that a K-bentonite type of mineral may be present, inasmuch as there is no shift of the x-ray peaks after glycol treatment.

34. Decatur-Dewey-Clarksville soils normally develop on limestone and dolomite and are kaolinite-rich throughout the profile (table 1; 7). The abundant illite in the soil in this sample undoubtedly has been derived from the illite in the parent shale.

35, 36, and 43. As in these soils developed on limestone and dolomite, published Hagerstown and Frederick profiles (table 1; 10, 14, 15) contain kaolinite and illite, but no montmorillonite—

a composition that is reflected in their low K_2O content and molecular SiO_2/Al_2O_3 ratios.

42. The matching differential thermal curves of this pair of samples (fig. 2) are unusual in that x-ray curves indicate the presence of kaolinite yet there is no sharp exothermic peak of kaolinite at 980° on the differential thermal curves.

45. Abundant illite in the non-marine Brunswick shale (Triassic) and in the fine-grained facies of the Stockton sandstone (Triassic) of New Jersey contradicts the generalization of some authors that kaolinite is the predominant clay mineral in non-marine sedimentary rocks.

CONCLUSION

The data assembled and discussed in this study support the conclusion that the clay mineral complex in soils developed on sedimentary rocks is commonly inherited from the parent material. Because illite is a very common constituent in the widespread mantle of sedimentary rocks, many of its derived soils contain an abundance of "relic" illite. In view of this relationship statements concerning the kind of climate favoring the formation of illite in soils must be based on examples containing authigenic illite, for inherited illite is present in soils in a wide variety of environmental conditions. The evidence adduced here also indicates that derived kaolinite is abundant in many of the soils developed on the Coastal Plain deposits of the southern regions. Montmorillonite is not commonly the predominant mineral in the clay fraction of soils, although it is associated with illite in both soils and sedimentary rocks.

A significant implication of these conclusions is that the alluvium eroded from widespread illite-bearing soils should contain considerable quantities of that clay mineral. Grim, Dietz, and Bradley (1949) have shown that illite is as abundant as kaolinite and montmorillonite in the clay fraction of the Little Colorado alluvium (table 1; 42) and in the clay fraction of near-shore deposits along the Pacific Coast (table 1; 43). In the sediments of the Gulf of California and in the deeper Pacific Ocean (table 1; 44, 45, 46), on the other hand, illite is most abundant and kaolinite least abundant. The authors believe that "the clay debris carried to the sea in many areas undoubtedly contains much degraded illite" (p. 1806) and have interpreted the increase in proportion of illite in off-shore marine sediments

as resulting from the slow loss of some of the kaolinite during diagenesis under marine conditions. The difference between the K_2O content of the Little Colorado alluvium and that of Gulf of California sediments (fig. 3; F, H) was thought to indicate that the degraded illite carried to the sea "would rapidly pick up potassium . . . because of its great avidity" for K_2O (p. 1806).

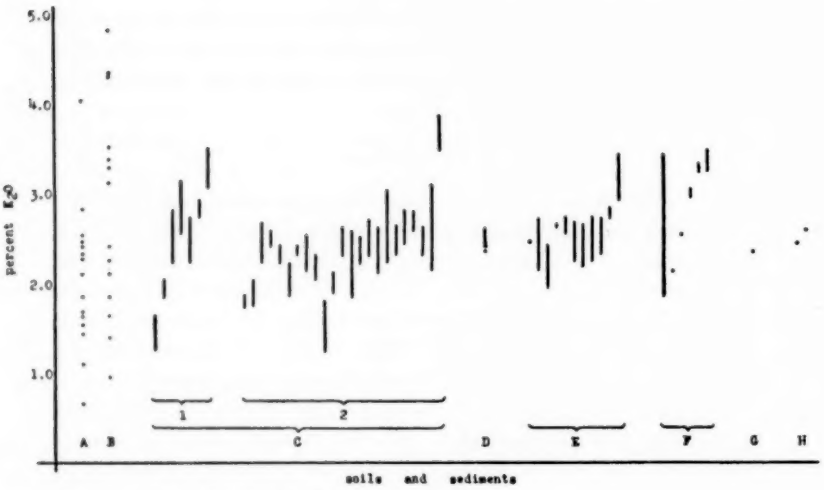

Fig. 3. K_2O Content of soils and sediments.
Dot—content of single analyzed sample.
Line—range of content in profiles with several analyzed samples.
A. A horizons of illite-bearing soils (tables 1 and 2).
B. Parent material or C horizons of illite-bearing soils (tables 1 and 2).
C. Alluvial soils of Mississippi drainage basin (Tech. Bull. 883).
1. Samples analyzed for clay mineral content (table 1; 47-53).
2. Other samples.
D. Gulf of Mexico sediments (Tech. Bull. 883).
E. Sediments of Pacific Ocean (Grim, *et al.*, 1949).
F. Sediments of Gulf of California (Grim, *et al.*, 1949).
G. Sediment of Mission Bay, California (Grim, *et al.*, 1949).
H. Little Colorado River alluvium, Arizona (Grim, *et al.*, 1949).

But comparison of their data with the K_2O content of soils, alluvium, and other marine deposits (fig. 3) does not suggest this conclusion. The amount of K_2O in the soils and alluvium of the Mississippi drainage basin, for example, shows no consistent variation with respect to that in the Gulf of Mexico deposits.

A detailed study of alluvial soils of the Mississippi River and its eastern and western tributaries (Holmes and Hearn, 1942) demonstrates conclusively that illite is the dominant clay mineral in the alluvium of this great drainage basin (table 1; 47-54). The analyses also reveal that the colloidal fractions of samples of sediments from the Gulf of Mexico (table 1; 55, 56) have essentially the same composition as the colloidal fractions of Mississippi delta alluvium, thus indicating that the composition of the alluvium contributed by the drainage system apparently is altered but little by the marine environment of the Gulf of Mexico.

From the source areas to depositional basins illite commonly is an abundant clay mineral. In many parts of the United States it is inherited in the soil from ancient sedimentary rocks, it is preserved in transportation and incorporated in modern sediments. But the inherited illite generally has been altered somewhat by weathering processes in the source area and as a result it has a variable K_2O content and molecular SiO_2/Al_2O_3 ratio. This alteration of illite should provide significant information about the effects of soil-forming processes, and it is obscured when illite minerals of differing composition are all identified as "illite" without more specific identification.

Throughout this discussion the similarity of the clay minerals in soil and parent sedimentary rock has been stressed. This emphasis is not intended to suggest that differences in composition between parent rock and derived soil or between different soils are not important or that authigenic clay minerals in soils are insignificant. Chemical analyses of profiles of Podzol soils, for example, show clearly that extensive fractionation of colloidal substances takes place in many soils developed on sedimentary materials. Similarly, the molecular SiO_2/Al_2O_3 ratios of the colloidal fractions of the Norfolk soils which contain inherited kaolinite range from 1.46 to 2.61. Inasmuch as this variation is clearly the result of different soil-forming conditions, it is important to an understanding of soil genesis. It should also be recalled that even though a soil contains some inherited illite, it may contain primary clay minerals produced by weathering of detrital minerals of the parent rock.

The kinds of materials and the relationship of the processes discussed here may result from environmental conditions peculiar to the present time, and may therefore be of limited use in

paleogeographic interpretations. Much of the soil that supplies the illite-rich alluvium of the Mississippi drainage basin is formed on widespread, illite-rich glacial drift and loess. Such glacial deposits were not common parent material during geologic history. In ancient times of more extensive seas, igneous and metamorphic rocks presumably were a greater proportion of the upland source areas than they are today, and on these crystalline rocks authigenic clay minerals developed in the soil.

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REVIEWS

Internal Constitution of the Earth; edited by BENO GUTENBERG. (Second revised edition of Volume VII of the "Physics of the Earth" series of the National Research Council.) Pp. 439; 88 tables, 43 diagrams. New York, 1951 (Dover Publications, Inc., $6.00).—The first edition of this well-known work, published in 1939, has been out of print for several years, and the revised edition will be welcomed by geologists and geophysicists. The revision, undertaken only a few years ago, has been accomplished with remarkable dispatch. Though the original plan has been retained, a number of chapters have been completely rewritten, many new references have been added, and other chapters have been added to in varying degrees. Two new authors, Benioff and Darling, have joined the original panel of nine; sixteen chapters, including an introduction and a summary by Gutenberg, deal with nearly every conceivable phase of the subject. The addition of new material has been accompanied by a considerable amount of pruning, and in spite of some thirty additional pages, the new book is thinner and more manageable than the old one. Produced by the Varitype process, it inevitably suffers in appearance and legibility by contrast with the printed first edition; in return, the price is low for a book of this size and complexity.

The only standard for comparison is the original edition, and by this test, the review of the literature in several fields is now somewhat less complete. Surprisingly, this is particularly true of seismology, various aspects of which are treated in several chapters. Although some references of so late a date as 1951 are included, there appears to be no discussion of the recent work on crustal structure by Hodgson, Ewing, Willmore, Gane, Tuve, and others, though a new table (49) has been inserted without comment. There seems to be no reference to any seismological paper by Jeffreys later than 1940, nor to the Jeffreys-Bullen tables, nor to the 1939 solution for the velocities in the interior of the Earth based on these tables. The study of local heat flow, a branch of geothermics which has developed since 1939, largely through the impetus given to this work by Bullard, is represented only by a short paragraph and table in chapter 6, and some brief comment in chapter 7. The great amount of information now available concerning the Earth's gravitational field receives only a few pages in chapter 12, with several inadequately reproduced figures of gravity anomalies.

In the new edition, chapter 2, by Jeffreys, on the "Origin of the solar system," begins where the corresponding chapter of the first edition left off, and brings the discussion up to 1948. Chapter 4,

by Adams, on "Elastic properties of materials of the Earth's crust," has been largely rewritten, as has chapter 7, by Gutenberg, on "The cooling of the Earth and the temperature in its interior." Chapter 13 on "Density, gravity, pressure and ellipticity in the interior of the Earth," has been revised by Lambert and Darling, and includes discussion of work up to 1948. Chapter 15, by Benioff and Gutenberg, is now entitled "Strain characteristics of the Earth's interior," and includes a summary of Benioff's recent investigations on the release of energy by earthquakes. Other chapters have additional references, new figures, and new tables. These samples of chapter headings suggest the encyclopedic nature of this volume, which even before revision remained an incomparable source of data and bibliography.

More discussion of some of the questions now under active investigation would have been welcome. For example, ideas of crustal structure which have prevailed for some twenty years are now undergoing revision as a consequence of new work in seismology. The nature of the evidence in this subject is left vague: there are tables of velocities, but no travel-time plots; pages of formulas, but no example of a seismogram. Concrete illustration of the complexity of the seismic record and of the difficulties of interpretation would go far toward clarifying some of the present points of controversy.

A book composed of contributions written independently by so many authors might be expected to have the defects associated with the obvious merits of this method; in fact, however, a very satisfactory balance has been achieved, and there are few important developments which escape mention. In a subject so diverse and so intrinsically difficult, systematic and continuous efforts toward synthesis are essential; the editor and other contributors to this volume deserve our warmest appreciation for their renewed labors in this direction. FRANCIS BIRCH

Families of Dicotyledons; by ALFRED GUNDERSON (with introductory chapters by Chester A. Arnold, Oswald Tippo, Theodor Just, Herbert F. Copeland, J. Herbert Taylor and W. H. Camp). Pp. xvii, 237; illustrated. Waltham, Mass., and New York, 1950 (Chronica Botanica Company: Stechert-Hafner, Inc., $4.75).—The author of this volume, Dr. Gunderson, was a member of the staff of the Brooklyn Botanic Garden for a period of thirty-two years, at first as an assistant in the herbarium and later as curator of plants, a position which he held until his retirement in 1945. His work afforded him abundant opportunity to study the dicotyle-

dons, and he developed a keen interest in the classification of this highly diversified group. The arrangement of the families and orders in the volume under discussion represents the results of much thought and study extending over a period of many years. The author has attempted to present a phylogenetic system based in part upon the classifications of other botanists, upon his own studies, and upon evidence afforded by researches in anatomy, paleobotany, cytology, and plant geography. He recognizes the limitations in devising a phylogenetic system on the basis of our present knowledge, however, and states: "The arrangement adopted is only tentative, not to be accepted but to be improved."

The book is copiously illustrated by artistic and accurate drawings from the pen of Miss Maud H. Purdy. There is a rather extensive bibliography, a good index, and the typography is pleasing. In brief introductory chapters experts deal with such topics as paleobotany, wood anatomy, cytotaxonomy, and plant geography.

Dr. Gunderson groups the dicotyledons into 42 orders and 240 families, in comparison with the 44 orders of Engler's Syllabus (edited by Diels, 1936), the 24 orders of Bessey (1915), and the 76 orders of Hutchinson (1925). Only seven of Gunderson's orders are monotypic as to family.

Whereas Hutchinson placed a great deal of emphasis upon the woody *vs.* the herbaceous habit, Gunderson states: "The change from a woody to an herbaceous stem must have occurred within many different families." In comparing Gunderson's system with those of Engler and of Bessey (the two which have received most support in America), a greater similarity with Bessey's system is perhaps apparent. In beginning his system with the Magnoliales (included in the Ranales by Bessey), Gunderson accepts the best evidence from several lines of study, and most students of phylogeny will agree that the most primitive dicotyledons are to be found in this group. He has avoided the pitfall of basing major groups upon a single difference such as archichlamydeae *vs.* metachlamydeae or cone flowers *vs.* cup flowers, realizing that such characters are quite artificial and have probably arisen independently more than once. His placement of the Caryophyllales near the Primulales, the Primulales near the Plantaginales, and the Umbellales near the Rubiales is suggestive of Bessey's arrangement, as is also the placement of the "Amentiferae" after the Hamamelidales (which Bessey included in the Rosales). There are many differences, however, such as Gunderson's treatment of the Violaceae and Cucurbitaceae as members of the Cistales, while Bessey considered that these two families are members of widely separated orders based upon the position of the ovary. Further differences are Gunderson's

placement far apart of his Campanulales and Asterales, and the same is true of Juglandiales and Fagales. This is made more striking when one examines his diagram on page 55.

We do not yet have evidence, if indeed we will ever have, which will allow us to see clearly the phylogeny of the dicotyledons. The relationships of some groups to each other seem obvious, but the affinities of others are very uncertain. Much in Gunderson's system is unique, and time will tell whether or not he has made sound choices in the composition and arrangement of his major groups. The community of taxonomists should welcome this volume which attempts to group the dicotyledons according to their relationships as determined from various lines of study. JOHN R. REEDER

Tensor Analysis, Theory and Applications; by I. S. SOKOLNIKOFF. Pp. ix, 335; 43 figs. New York and London, 1951 (John Wiley & Sons, Inc., and Chapman & Hall, Ltd., $6.00).—This book, which is the latest addition to the Wiley Applied Mathematics Series, has many excellent features, but the title is somewhat misleading. In brief, it is long on theory and short on applications; the latter being little more than a bare introduction to the subjects with only the most general indications of how the structure may be practically applied.

With this reservation, the book is an excellent treatise on tensor analysis. The author has gone to great pains to abstract those portions of modern higher algebra which have been found necessary to provide a mathematical basis for the field of applied mathematics known as tensor analysis. As such it provides the theorems which are pertinent, in a logical development which moves slowly enough for the student without the special training of the professional mathematician.

The first chapter on linear vector spaces and matrices might well be required reading for all students of quantum mechanics at some time during their introductory course. The second chapter on tensor theory is also excellent and completely self contained. The third applies the results of the previous two to the problems of differential geometry. The remainder of the book is devoted to introductions to analytic mechanics, relativistic mechanics, and mechanics of continuous media.

This text will appeal especially to the relatively mature student who is reasonably familiar with one or more of the fields of application of tensor analysis and wishes to systematize his understanding of the subject while placing it upon a firm mathematical foundation. D. E. HARRISON, JR.

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